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Astronomical Theory of Paleoclimate: A New Concept

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Abstract—Essentials of Milankovitch theory are considered in order to reveal its consistency with paleoclimatic and astronomical data obtained during the last 30 years. Numerical data calculated by Milankovitch are shown to be inappropriate for paleoclimatic interpretations due to a series of outlined reasons. A new concept of astronomical theory of paleoclimate is suggested. Its essence consists in the new, as compared to the previous one, substantiation of climatic influence of orbital elements variations, primarily of the eccentricity fluctuations during the last million years, whose impact is shown to be exclusively significant and decisive. The insolation-climatic diagram plotted for the last million years is well compatible with empirical paleoclimatic data (oxygen isotope records).

Key words: *astronomical theory, Milankovitch theory, paleoclimate, insolation, eccentricity, obliquity, precession, oxygen-isotope records, solar activity*

INTRODUCTION

The astronomical theory of paleoclimate (ATP), as the theoretical basis for paleogeographic, paleoclimatic, and chronostratigraphic analyses, is of a great significance for investigating the Quaternary period. The ATP was created to reveal relationships of the recurrent Pleistocene ice ages with the changes in insolation of the Earth in response to variations of the Earth's orbital elements. The ATP development began about 150 years ago. Cyclical changes of the Earth's orbital elements, e.g., of the eccentricity e with a period of about 100000 years and of the ε angle between the equatorial and ecliptic planes (termed further as the obliquity angle) with a period of 41000 years, have been recognized already in the 19th century along with the 23000-year cycle of precession of the equinoxes, the last cycle thought to be related basically to the Earth's axis precession.

The climatic consequences of precession are as follows: due to the ecliptic of the Earth's orbit, the climate of northern and southern hemispheres should be characterized alternately by a long cool summer and short mild winter, or by a short hot summer and long cold winter. In the northern hemisphere, the first climate type is in the time period when the Earth is in aphelion (at the greatest distance from the Sun) and this hemisphere faces the Sun. After 11500 years (the precession half-cycle), the rotation axis of the Earth passing the aphelion points to the Sun with its southern end, and the first climate type shall be dominant now in the southern hemisphere, whereas the other climate must be expected in the northern one. English mechanic and mathematician J. Croll (1875) believed that glaciations in the northern hemisphere are the climatic results of long and cold winters.

All three orbital elements of the Earth have been considered most comprehensively by Yugoslav scientist M. Milankovitch, who was the first to calculate quantitative changes of insolation that arrived at the Earth, more precisely, at the top of the atmosphere during the last 600000 years. Using these data, he evaluated the temperature fluctuations at the planet surface. According to Milankovitch (1939), the obliquity corrects first of all the latitudinal distribution of insolation. These relative changes are most significant within high latitudes, being less notable in low latitudes, where they are of the opposite sign. The lesser the ε angle, the lower the insolation in high latitudes, and this, according to Milankovitch, should increase the glaciation area around the planet poles. The positive feedback because of the greater albedo shall be favorable for the further growth of ice sheets, and the ε decrease thus predisposes the planet to development of glaciation. Considering the precession influence, Milankovitch surmised, in contrast to Croll, that glaciations develop in response to long cool summers and short mild winters, but not in periods of long severe winters and short hot summers. *It is necessary to note here that neither variations of obliquity nor precession are able to change the annual amount of solar radiation arriving in sum at the Earth; they are responsible only for the seasonal or latitudinal redistribution of insolation. In contrast, variations of eccentricity regulate the annual value, or more precisely, the power of insolation.* As shown by Milankovitch (1939), the insolation variations are proportional

to $\frac{1}{\sqrt{1-e^2}}$, where e is eccentricity, and their relative peak difference for the last 600000 years was only 0.11% of the total solar radiation coming to the Earth. This quantity is much less than the relative insolation changes controlled by variations of the axial tilt ε and precession. Accordingly, he considered the eccentricity alteration as having no individual contribution to insolation changes, taking this parameter into consideration, when the solar energy flux to the upper atmosphere boundary was under evaluation, only as a parameter modulating the precession-related changes, because the insolation amplitude dependent on the last factor must increase, given the greater e magnitude. Variations of the solar radiation (ΔQ) increment at the upper atmosphere boundary within different latitudes are described for the summer (ΔQ_s) and winter (ΔQ_w) semiannual periods of northern hemisphere by the following two equations (Milankovitch, 1939):

$$\Delta Q_s = \Delta W_s \Delta\varepsilon - m\Delta(e\sin\omega), \quad (1)$$

$$\Delta Q_w = \Delta W_w \Delta\varepsilon + m\Delta(e\sin\omega), \quad (2)$$

where ΔW is the insolation difference dependent on the given latitude, when the axial tilt increment $\Delta\varepsilon$ is 1°; m is the coefficient dependent on the latitude; and ω is the so-called longitude of perihelion. In the case of the southern hemisphere, the above formulas must replace each other, because the semiannual insolation changes dependent on precession are in antiphase for opposite hemispheres in contrast to the monophase insolation changes in both hemispheres, which are controlled by obliquity. The global manifestation of the precession cycle is thought to be dependent on the dissimilarity of hemispheres, and the northern one is a common place of scientific interest, because its ice sheets had the decisive role in development of the global Pleistocene glaciations. As one can see from the quoted formulas, the *annual* precession-dependent insolation increment at any given latitude (the sum of second terms in the right part of both formulae) is equal to zero, and Milankovitch calculated the semiannual insolation sums in order to evaluate the contribution of precession. The equation (1) was used to plot the famous insolation diagram by Milankovitch, who calculated data for 65° N, the latitude considered to be most sensitive to the insolation changes (Fig. 1). Naturally, this diagram does not show distinct periods related to the eccentricity alteration, which should be 100000 years long. According to calculations by Milankovitch, there were four glaciations on the Earth during the last 600000 years, and this conclusion perfectly fitted the well-known climatic scheme by Penck and Bruckner, where they distinguished the Würm, Riss, Mindel, and Günz glacial epochs. This direct correlation between the Milankovitch insolation diagram and principal climatostratigraphic subdivisions was even used to date the latter and to create the Pleistocene chronostratigraphic scale. However, the later oxygen-isotope (OI) data on oceanic bottom sediments (Shackleton and Opdyke, 1973, 1976; Hays *et al.*, 1976) modified this concept, and inside the Brunhes Chron, there were established eight glaciations (Bowen, 1981). The OI data were inconsistent not only with the Milankovitch plot, but also with other insolation diagrams more precisely substantiated by Sharaf and Budnikova (1967, 1969), Berger (1978a, 1978b, 1988) and some other researchers. Despite this contradiction to the Milankovitch theory, the spectral analysis of OI records (Hays *et al.*, 1976) revealed climatic cycles, the duration of which was close to 100, 41, and 23 thousand years, i.e., to variation periods of the Earth's orbital elements. These cycles of OI records in bottom sediments were used to revise age of the Matuyama–Brunhes reversal. The reversal was estimated to be about 780 thousand years old (Johnson, 1982; Shackleton *et al.*, 1990) that was proved later by radiometric dating (Spell Terry and McDougall, 1992; Shackleton, 1995). The Milankovitch theory thus found wide acceptance among scholars. In our country, Krasnov (1974) was first to pay serious attention to the correlation between its theoretical postulates and real geological data; he also initiated new works aimed to calculate more precisely the variations of orbital elements.

FORMULATION OF THE PROBLEM AND OBJECTIVES

Despite its general success, the Milankovitch theory arrived at a paradox. On the one hand, all three periods of the orbital elements variations are manifested in climatic changes, but on the other, the direct comparison of climatic records with theoretical insolation variations calculated by Milankovitch showed their discordance. This discordance is illogical if we believe that the orbit-dependent variability of insolation controls the climatic changes. It should be mentioned as well that the 100000-year cycle distinctly established in climatic changes during the last million years is taken to be insignificant in the Milankovitch theory. The contradictions discredit the theory, which is accepted as a theoretical basis for the universal usage of OI records as an equivalent of the climatochronostratigraphic scale of the Pleistocene. They also suggest that, in addition to the orbital elements variations, there could be some other factors considerably influencing the climate. In particular, the last point is evident from the fact that the periods of orbital elements variations and respective climatic changes were quite regular during the Mesozoic and Cenozoic (Zhemchuzhnikov, 1963; Berger, 1988; Chumakov, 1995), although the most significant ice sheets developed on the planet during the last million years only. The Milankovitch theory needs to be further refined, as its imperfections decelerate development of the theory proper and entangle the correct understanding of Quaternary climatic changes.

Thus, considering the theoretical consequences and experimental data, the OI records in particular (Figs. 1, 2),

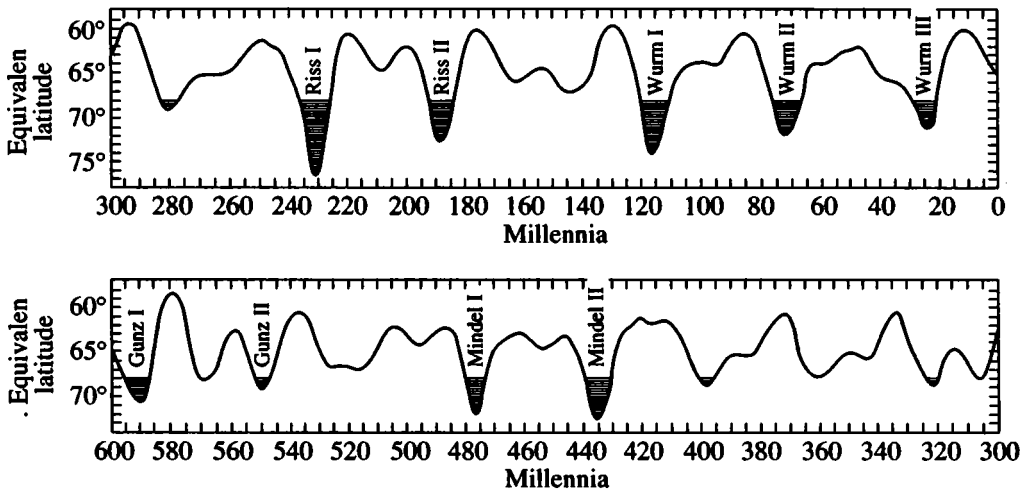

Fig. 1. Secular variations of insolation intensity in a form of latitudinal equivalents (Milankovitch, 1939). The insolation minimums and maximums correspond to cooling and warming stages respectively; shaded areas mark glaciations considered by Milankovitch.

we may conclude that principal weak points of the Milankovitch theory are as follows. (1) The amount and ages of glaciations, as they are depicted in the insolation diagram by Milankovitch and actually recorded by OI data within the Brunhes Chron, are at variance. Moreover, analyzing the more precise insolation diagram by Berger (Fig. 2b), one can see not only this discordance, but also the total absence of criteria for distinguishing the glaciation periods using diagrams of this kind. (2) Glacial periods correspond to intervals of the decreased eccentricity in the diagram (Fig. 2) that have no explanation in the theory. (3) The 100000-year periods prevail in all known $\delta^{18}O$ variation curves for the last one million years (Hays *et al.*, 1976; Imbrie *et al.*, 1984, 1993; Shackleton, 1995). It is precisely these periods approaching the eccentricity variation cycles that are underestimated by Milankovitch and his disciples. (4) The theory is based on the presumption that the power of radiated solar energy was constant during the geological time, but neither this nor the opposite statement is seriously substantiated.

ANALYSIS OF THE THEORY ESSENTIALS FROM THE VIEWPOINT OF EMPIRICAL DATA

Recent data, in contrast to opinions by Milankovitch, suggest variable, rather than steady solar luminosity and the climatic significance of these variations. The satellite observations (Willson and Hudson, 1991; Sklyarov, 1994; Nesme-Ribes *et al.*, 1996; Chistyakov, 1997) point to the cyclical variations of the solar constant S_0, which are as great as 0.1% being in phase with the 11-year cycles of the solar activity (SA). The S_0 maximum corresponds to the Wolf number maximum. There are noted as well the longer periods of SA variations. These are the SA reductions named as Wolf (of ≅1281–1347), Sporer (of ≅1400–1510), and Maunder (of 1645–1715) minimums and the preceding medieval SA maximum (Eddy, 1978; Krakovetskii *et al.*, 1985; Vitinskii *et al.*, 1986; Ivanov-Kholodnyi and Chertoprud, 1990). The last event was responsible for a minor climatic optimum of the 10–12th centuries, whereas the SA minimums were related to the global "Little Ice Age" of the 14–18th centuries (Eddy, 1978; Hays, 1980; Klimanov, 1994, 1997; Damon, 1980; Nesme-Ribes *et al.*, 1996; Chistyakov, 1997; Gallardo, 1997). In many publications (Eigenson, 1963; Fairbridge, 1966; Eddy, 1978; Hays, 1980; McCormak and Seliga, 1982; Mitchell *et al.*, 1982; Chistyakov, 1977; and others), the SA variations used to be considered as controlling climatic changes, but this idea is not always well substantiated (Monin, 1969; Pittock, 1982). It is possible to assume that the regional character and instability of climatic events during the 11-year cycles depend not only on a weak changes of summary radiation, but also on *the relatively short-term influence* of this factor (11 years) on the highly energetic and inertial climatic system of the Earth. This particular factor appears to be insufficient for global changes. From this standpoint, the long-term SA variation periods and related climatic changes are much more significant. The correlation between the SA and climate seems to be substantiated for the Maunder minimum, because the Wolf numbers and climatic parameters of that time are partially estimated on the basis of tool observations, historical information, and dendrochronological data (Eddy, 1978; Dergachev, 1994; Nesme-Ribes *et al.*, 1996; Chistyakov, 1997; Baliunas *et al.*, 1997). The established relation of the Maunder minimum with the peak phase of the "Little Ice Age" means that the long-term (longer than 11 years) SA decrease may cause global climatic cooling. Accordingly, it seems logical to assume that much longer periods of insolation variations of a similar magnitude (~0.1%), which are connected with the eccentricity difference, could be all the more responsible for climatic changes. The assumable personal sig-

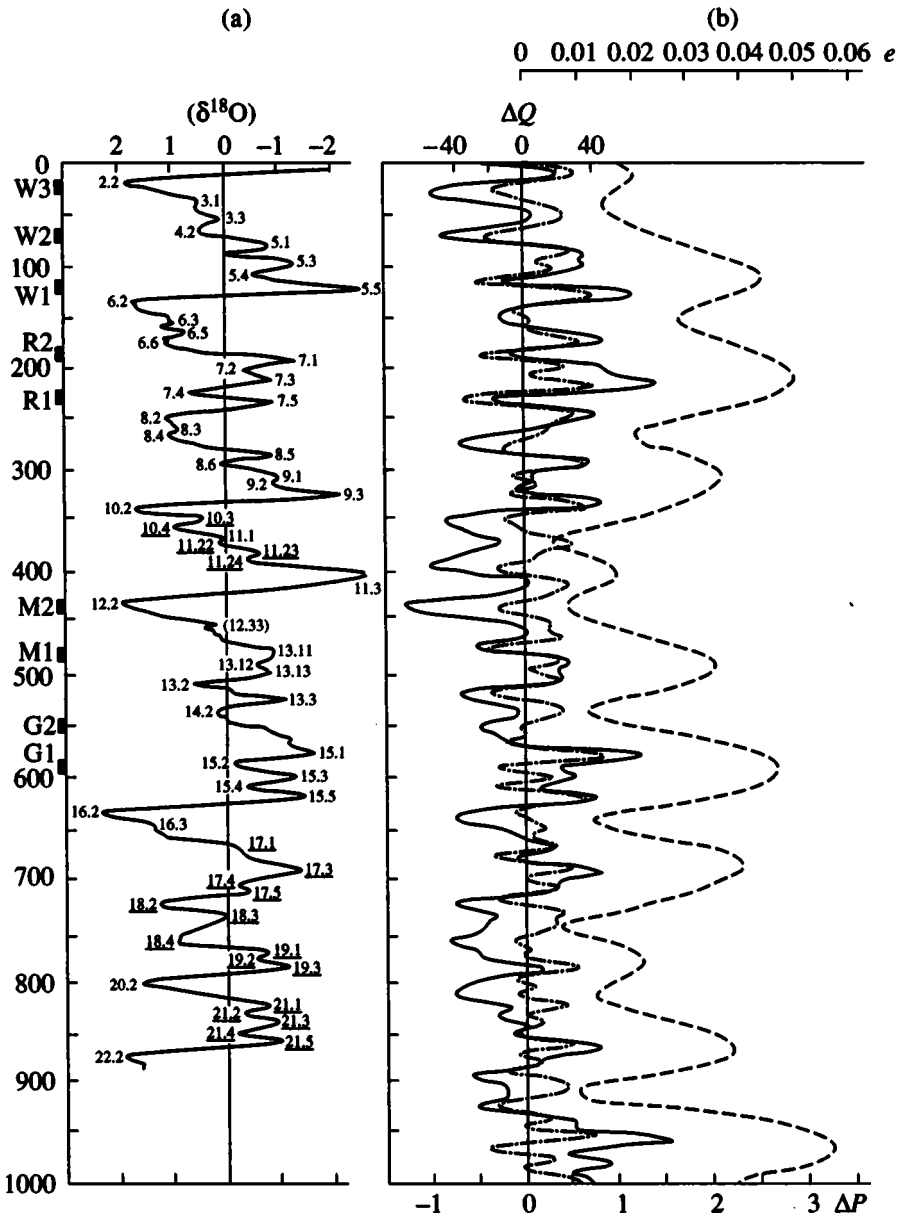

Fig. 2. Comparison of the oxygen isotope scale with the insolation-climatic diagram. (a) Oxygen isotope scale by Bassinot *et al.* (1994): glacial (cooling) and interglacial (warming) stages and substages are denoted at the curve by even and odd numbers respectively; shaded rectangles at the left side show the ages of the glacial stages Wurm (W 1, 2, 3), Riss (R 1, 2), Mindel (M 1, 2), and Gunz (G 1, 2) after Milankovitch. (b) The solid line represents the insolation-climatic diagram plotted in this work, the dashed line shows variations of the eccentricity e (after Berger, 1978), and the dashed-dotted line illustrates the solar radiation deviation ΔQ (in arbitrary units) relative to the caloric summer of 1950 in the northern hemisphere (after Berger, 1978). Arbitrary unites ΔP of the insolation-climatic diagram are proportional to the possibility of glaciation (negative values) or interglaciations (positive values). Ordinate shows time in thousand years before present.

nificance of eccentricity for the climate finds support from experimental data. First of all, this is the manifestation and even predominance of the 100000-year periods of climatic changes during the last million years. Second, there is a distinct coincidence of ice ages (the even OI stages) with the intervals of the eccentricity minimums (Hays *et al.*, 1976; Imbrie *et al.*, 1984, 1993; Fig. 2). The minimum value of solar radiation arriving at the planet corresponds exactly to the lowest e magnitude.

An obvious argument against the above assumption—that the eccentricity is a factor of a separate climatic significance—consists of the theoretical appraisal of Milankovitch, according to which the relative insolation variations dependent on the precession and ε changes are more than one order of magnitude

greater than those dependent on eccentricity. Accordingly, the contribution of the last factor can hardly be manifested.

The prevalence of 100000-year climatic cycles during the Pleistocene has been studied for more than 20 years by many researchers outside this country, for instance, by Hays *et al.* (1976), Imbrie *et al.* (1984, 1993), Berger (1988), and Berger and Loutre (1992). Imbrie *et al.* (1993) considered various, so-called nonlinear mechanisms of the eccentricity signal amplification, in response to which it could be manifested against the background of more intense, as Milankovitch believed, insolation changes related to obliquity and precession. In our opinion, there is an alternative approach based on the presumption that the *real* influence of obliquity and precession on the planet surface heating by the Sun and related climatic changes could be less than that theoretically calculated by Milankovitch. Below, we present our attempt to substantiate the possibility of such a decrease in terms of physics.

Evaluating changes in the Earth's surface temperature, Milankovitch used the insolation parameters theoretically calculated for the upper atmospheric boundary. According to his theory, the atmosphere was considered as immobile and the Earth's surface as homogeneous. In other words, the main climate-forming factors of the Earth, namely the coexistence of global ocean and continents, as well as the dynamic interaction and circulation of atmosphere and hydrosphere, were ignored. All these features considerably complicate the global climatic model, although some consequences of the atmosphere–ocean dynamics are rather clear. As known, the circulation of atmosphere and hydrosphere evens the temperature gradients between different regions of the planet. As a consequence, this gradient between the equator and poles is less than that theoretically derived, and Milankovitch (1939) personally noted this situation. Thus, the temperature distribution over the Earth is basically under the control of two factors: the solar radiation distribution dependent on astronomical parameters and the terrestrial processes of heat exchange, which act in this case as the negative feedback causing a decrease of the thermal gradient between poles and equator. No doubt, this negative feedback related to the circulation of atmosphere and hydrosphere must be active, as well in the case when the solar heat distribution over the planet surface is governed by changes in the Earth's axial tilt.

Since the obliquity increase is favorable for the greater heat flux into high latitudes and, consequently, for the decrease of thermal gradient between poles and equator, this event and circulation of the atmosphere and hydrosphere should act together decreasing the ice volumes in high latitudes and thus leading to the climatic warming. In contrast, the lesser axial tilt results in cooling within these zones and, as Milankovitch believed, over the globe. Accordingly, the atmosphere and hydrosphere circulation must be more intense during the cooling periods induced by decrease of the axial angle ε, because the temperature gradient between the poles and the equator grows. Such an intensification of atmospheric circulation during the ice ages has been noted in many works (Beget and Hawkins, 1979; Hovan, 1979; Imbrie *et al.*, 1993; and others). This is an argument in favor of the climatic contribution of ε parameter suggested by Milankovitch.

Thus, the atmosphere and hydrosphere circulation must act against the cooling processes in high latitudes, which are related to the axial tilt decrease. In other words, it must result in a real reduction of the climatic contribution of obliquity variations as compared to that calculated by Milankovitch for the top of the atmosphere. We have no numerical values describing to what extent the solar heat amplitude dependent on obliquity actually changes at the Earth's surface relative to the amplitude theoretically derived by Milankovitch on the basis of his simplified model. Taking into account that the 100000-year climatic cycles prevail over the 41000-year cycles, we may, however, assume a considerable amplitude decrease.

Let us now consider the precession. As the annual insolation changes dependent on precession are equal to zero for any given latitude, Milankovitch had to calculate the semiannual insolation sums in order to distinguish the climatic contribution of the precession. This approach appears to be artificial in terms of its subsequent climatic interpretation. Actually, it seems rather odd to think that the insolation has a climatic influence during one half of the year and is inert during the other. Moreover, the semiannual insolation variations related to precession are in antiphase, and the summer insolation increase antecedes its decrease during the winter, and vice versa. (It would be strange to expect the apple crop in autumn if the day temperature during summer were to become positive after winter, but the night temperature remains negative as before.) *The aforesaid means that the semiannual insolation magnitudes cannot be used for the evaluation of summary changes in solar radiation coming to the Earth during the period exceeding one year.*

There are some other arguments in favor of our conclusion. In our opinion, the great energy and inertness of the Earth's climatic system imply an averaging (and accumulation) of small insolation changes and their summary impact on the planetary climate for a period considerably greater than half a year. Consequently, in evaluating insolation changes for a given geological time span, one must take into account the summary annual, but not semiannual insolation, since it is precisely the former that characterizes the main seasonal climatic cycle of the planet. More than 150 years ago, Humboldt (1848) asserted the importance of considering the annual, rather than semiannual temperatures (Imbrie and Imbrie, 1988), and this statement actually means that the precession-dependent contribution to insolation is equal to zero (see equations 1, 2), i.e., the real climatic influence of precession, as compared to

that accepted in theoretical considerations by Milankovitch, narrows down. Nevertheless, it would be incorrect to think that the precession does not influence the climate. Its influence, having no direct connection with quantitative insolation changes, realizes itself by means of another mechanism. For instance, a transition from a long cool summer and short mild winter to a long cold winter and brief hot summer, which is controlled by precession, can be correlated with humidity changes in respective hemispheres. Such a presumption agrees with the fact that the cycles of lake level fluctuations during the Late Triassic, which presumably reflect the variable humidity, are 23000 years long (Olsen, 1984).

Thus, there are real arguments, factors, and mechanisms that minimize, relative to the contribution theoretically inferred by Milankovitch, the climatic influence of insolation variations ε dependent on obliquity and precession. Naturally, this situation highlights the significance of the eccentricity factor. Milankovitch well understood the importance of the Earth's atmosphere and hydrosphere for dynamic redistribution of solar energy arriving at the upper atmospheric boundary. He was stressing that his numerical results are correct for the immobile atmosphere and homogeneous Earth's surface. The precise mathematical analysis of dynamic influence of atmosphere and hydrosphere on the solar heat distribution over the atmosphere top surface is a difficult objective even at present. The same seems to be true with respect to modeling the climatic changes related to the precession as we understand them in this work.

PLOTTING MODE OF INSOLATION-CLIMATIC DIAGRAM

As we have seen above, the real climatic significance of orbital element variations must be other than that suggested in Milankovitch's theory. By solving the problem, one should take into account the specific influence of positive and negative feedback acting in terrestrial conditions and transforming the solar signal dependent on variations of three orbital elements into the climatic changes. The precise mathematical solution of the problem is a prerogative of specialists in climatic simulations.

Thus, Milankovitch's idea to plot the diagram of secular climatic changes on a strict mathematical basis can be realized only with considerable difficulty. In our opinion, there is another way of solving the problem with due account for the empirical data. As compared to the time of Milankovitch, we now have many more possibilities for evaluating the climatic effect of orbital elements variation using the diverse empirical data on paleoclimatic oscillations.

Principal for our consideration is the above presumption that the axial tilt and eccentricity decrease leads to climatic cooling so that it will be of global significance, when the summary magnitude of both orbital parameters will be at the minimum. If the precession is also included in the examination, then, in accordance with the mechanism of its climatic influence, we should consider only one, namely, northern hemisphere. In order to get the uniformity of the climatic changes in coordination with all orbital elements variations, it is necessary to plot a graph of precession variation where minimums would correspond to the cooling events. In this case, the term $\Delta(e \sin \omega)$ of the equation (1) must be taken with the negative sign, and then the summary minimum of all three orbital parameters would indicate the glaciation in the northern hemisphere. In other words, if we know the coordinated variations of three orbital parameters with time, we can define the most probable time intervals, when glaciations may develop (assuming, of course, the decisive climatic significance of insolation variations dependent on parameters of orbital elements). Since changes in the orbital parameters are different in their physical nature and magnitude, they must be reduced in a uniform mode. The common practice in this case is normalization, and we did it in a most simple way. Initially, the orbital elements variations have been calculated relative to their arithmetic mean for a given time interval (million years). In this case, the obtained values were positive when they exceeded the mean parameter, and negative, when they were lesser than the latter. Afterward, the obtained values were divided by the maximum parameter recorded within the interval of one million years. Therefore, the diagrams or a sequence of magnitudes characterizing the orbital element variation will be within the interval ±1. According to our interpretation, the negative peaks are possible indicators of glacial periods, whereas the positive ones correspond to interglacials. The cumulative variations of all orbital elements can be depicted by means of a superposition of three plots, which are now of the same scale, but first we must take into account that different orbital elements are of unequal climatic significance.

According to results of harmonic analysis of OI records, the most significant climatic changes of the last million years are typical of the 100000-year cycles; the 41000-year cycles are second in importance, and the least amplitude of changes is characteristic of periods 23000 years long. Judging from these data, we may conclude that the main orbital parameter responsible for climatic changes within the time interval under consideration is eccentricity; obliquity is next in importance, and precession is the last one. It seems reasonable to think that the relative climatic contribution of different orbital elements is proportional to amplitudes of climatic changes characteristic of above periods. Using climatic diagrams published by Imbrie *et al.* (1993), one may calculate that ratios between these amplitudes for the periods of 100, 41, and 23 thousand years vary from 1 : 0.52 : 0.38 to 1 : 0.86 : 0.45. According to these data, the proportion 1 : 0.71 : 0.43 that was taken quite conditionally by one of us (Bol'shakov, 1998) to characterize the relative climatic contribution

of eccentricity, obliquity, and precession, appears to be realistic.

Hence, we have to give a different weight to the three diagrams under consideration introducing, for instance, the respective "climatic" coefficients for Y-axis of each plot. Naturally, the greatest coefficient (the highest amplitude of changes) should correspond to eccentricity variations, and the least one to precession. Taking into account the data by Imbrie *et al.* (1993), the coefficients 1 : 0.7 : 0.4 can be considered as realistic for the eccentricity, obliquity, and precession, respectively. Their precise values can hardly be estimated at present. For instance, the aforementioned work by Imbrie and his colleagues shows that the influence of climatic cycles of different duration varies with time and is at variance when estimated on the basis of different climatic indicators, such as the OI records, paleotemperature analysis, carbon isotope variations in carbonates, etc.

Thus, our diagram depicting the cumulative variations of orbital elements is obtained by means of the superposition of all three plots corrected in accordance with the accepted "climatic" coefficients (Fig. 2b). Numerical data on the orbital elements variations are adopted from the work by Berger and Loutre (1991). According to its intent and essence, the diagram can be termed as the "insolation-climatic" (IC) graph, as it shows the relative probability ΔP of advent of glacial (negative peaks) and interglacial (positive peaks) stages. The term "relative probability" (i.e., greater or lesser) points to the qualitative chance to detect climatic events in the diagram.

DISCUSSION

The IC diagram shows a much better correlation with OI records in contrast to insolation diagrams by Milankovitch and his followers (Figs. 1, 2). In addition, it should be noted that Berger (1978b) also considered the paleoclimatic interpretation of insolation diagrams as being unsound. Using the IC diagram, one can detect almost all OI stages for the last million years. The most interesting is the positive peak in the IC diagram versus the time interval of about 770–790 ka, which corresponds to the 19th OI stage of warming. It is precisely this stage that marks the Matuyama–Brunhes reversal, whose age, according to recent determinations, falls within the above time interval. All even OI stages, except the sixth one, correspond to the *most significant minimums in the IC diagram*, which are, in their absolute magnitude, greater than or identical to $\Delta P = 0.5$ of the adopted conventional scale. The age position of the minimums somewhat predates, as it should, the respective maximums of glaciations. There are also some examples of remarkable visual similarity. These are, for instance, the double minimum of the OI stage 18, the triple maximum of the stage 15, and a deep characteristic minimum within the first half of the "warm" stage 7. As one can see, glacial stages correspond to the intervals of lowered eccentricity values. Bearing in mind the plotting mode of the IC diagram, we may conclude that it reflects the variation cycles of all three orbital elements, including those 100000 years long. Hence, the diagram eliminates all the aforementioned contradictions between the astronomical theory and experimental data.

Nevertheless, we should also mention some differences between the IC diagram and OI scale. The most notable is the disproportion in amplitudes between the peaks of 11th and 6th OI stages and corresponding maximum and minimum in the IC diagram. We consider this discordance as a consequence of several factors.

(1) Postulates used as basic in the data processing are qualitative in many respects. On the one hand, this is unavoidable at the initial stage of investigation, but on the other, the generalized approach has its own positive aspects. The IC diagram reflects planetary climatic changes and thus better corresponds to the global extent of glacial and interglacial periods of the Pleistocene in contrast to insolation diagrams by Milankovitch and his adherents, which characterize the separate latitudinal zones. It is noteworthy as well that the insolation diagrams for high and low latitudes are in antiphase within the considerable time interval, and this triggers various speculations. The soundness of our IC diagram naturally depends on the precision and actuality of the calculated orbital element variations and related insolation changes. Concerning this problem, we can only note that two IC diagrams based on numerical data by Sharaf and Budnikova (1969) and by Berger and Loutre (1991) are similar in many aspects, and we used the latter set of data, since it is more precise.

(2) The inadequate interpretation of IC diagrams depends on various circumstances, as it can be seen, in particular, from comparison of different OI curves variable in shape and depicting OI stages of inconsistent duration and age (Bol'shakov, 1998).

(3) According to available data, the idea that variations of the solar constant may influence the Earth's climate has at least the right to exist. However, it is impossible at present to prove this influence for the geological past, because we have no chance of determining the respective time-dependent variations of S_0.

(4) Significance of terrestrial factors for climatic oscillations (changes in the atmosphere composition, in albedo of the Earth's surface, etc.) has been considered to some extent earlier. It is evident in addition from analysis of OI records (Ruddiman *et al.*, 1986; Imbrie *et al.*, 1993), according to which climatic cycles with the period of 41000 years dominated 1.8–1.2 Ma ago, whereas the 100000-year periods prevailed during the last million years. It is interesting to note that the most significant $\delta^{18}O$ excursions of the last period mark the most intense cooling events detected from OI records (Shackleton, 1995). Some researchers believe that changing rhythms and amplitudes of climatic oscilla-

tions are related to the ice volume growth in response to the global temperature decrease that begun in the Eocene–Miocene time. One million years ago, the global ice mass might have approached a certain critical value sufficient for development of auto-oscillations in the system ocean–atmosphere–cryosphere–lithosphere. According to some theoretical considerations (Sergin and Sergin, 1969; Oerlemans, 1980; Budyko, 1981; Imbrie *et al.*, 1993), a period of these oscillations can be 60–120 thousands years long. Taking into account that the IC diagram is well correlative with the OI records, the oscillations should not be free, but forced to a certain extent and controlled by insolation variations dependent on orbital parameters.

In our opinion, the mechanism of *forced auto-oscillations* sustaining the 100000-year-long global climatic cycles can bear some explanations for the discrepancy between the IC diagram and OI records that is noted above. However, the new ATP concept is at the initial stage of development, and there could be some other reasons responsible for discordance of theoretical and experimental data.

CONCLUSION

Summarizing the results of the study, it is necessary to emphasize the following facts. The titanic work by Milankovitch, who elaborated the mathematical substantiation of the ATP, was a great contribution to fundamentals of the theory. In our opinion, however, Milankovitch was off on two points when interpreting the obtained numerical values in terms of paleoclimatic fluctuations. First, the *global* climatic changes dependent on the obliquity angle were considered on the basis of insolation variations defined for a single favored latitude. Yet the concept of global climatic oscillations means the tight interrelation between changes in climate over the whole planet, and this cannot be adequately understood, if we consider the insolation variations at one latitude only, so much more that they are opposite in the high and low latitudes. Therefore, it is more correct to calculate the summary insolation variations for all latitudes and their transformation into climatic changes. (Ignoring this, one actually disregards or destroys the planetary climatic connections.) The calculation of this kind for the top boundary of the atmosphere would result in the zero climatic effect, because variations of the ε parameter does not change the annual insolation of the Earth as whole. Nevertheless, the climatic effect of the obliquity angle is clearly evident from a comparison of the IC diagram and OI records for the last 120000 years, and Milankovitch correctly understood its mechanism (the decrease in ε parameter triggers cooling). The above consideration implies that the global climatic effect of the obliquity angle must be predominantly nourished by the positive feedback primarily intensifying the insolation variations in high latitudes (for instance, because of the changeable albedo in response to changes in the ice sheet areas). Consequently, it is necessary to evaluate the insolation variations within different latitudinal belts and their climatic effect under the real terrestrial conditions. Apparently, the model must take into account the atmosphere and hydrosphere circulation, albedo, concentration of carbon dioxide and dust in the atmosphere, and other factors specific for various latitudes and sustaining the climatic effect of ε variations.

The next weakness in the Milankovitch theory is related to the dismembered time succession of the accumulating insolation influence, because the precession-related impact on global climatic changes, whose period is about 23000 years, was inferred using the semiannual, but not annual, insolation sums, although exactly the latter determine the basic seasonal climatic cycle of the planet. The latter, as was known even before Milankovitch, are equal to zero for any latitude, and the real climatic contribution of precession could not be as high as he believed. Considering all the data discussed above, we should presume the greater significance of eccentricity variations. Consequently, the numerical values calculated by Milankovitch do not correspond to the real climatic influence of orbital elements and cannot be used in interpretation of climatic changes.

Thus, analyzing basic postulates and conclusions of Milankovitch theory along with empirical data (OI records), we may conclude that the climatic influence of the obliquity angle and precession is not as great as suggested in the theory, and the eccentricity variations make their own significant contribution to climatic changes. Our appraisals of the climatic influence of the Earth's orbital elements differ from those by Milankovitch. Together with concrete physical mechanisms approving their reliability, these appraisals represent the essence of the new ATP concept. Having no precise determination of the individual contribution of every orbital element variation to the climatic changes, we used the empirical data to evaluate the relative significance of particular factors. It is important as well that we not only accepted the climatic influence of orbital elements as proportional to amplitudes of respective climatic cycles with different time spans, but also *demonstrated qualitative physical preconditions for the adopted ratio* (Bol'shakov, 1998). When plotting without this physical substantiation, the ETP graph of integrated orbital element variations, Imbrie *et al.* (1984) failed to introduce the "climatic" coefficients like those described above. Using the new concept, numerical data by Berger and Loutre (1991), and analytical data by Imbrie *et al.* (1993), we plotted the IC diagram reflecting the relative probability of glacial and interglacial stages during the last million years. The procedure was based on the very general and understandable climatic influence of the eccentricity and obliquity angle of the Earth's orbit. (The problem of the precession influence is much more complicated, and we followed here Milankovitch's idea that glaciations in a given hemisphere are favored by climates with a long

cool summer and short soft winter. In our interpretation, this may correspond to the climate of a greater humidity degree.) Such an approach is rather qualitative but has its own advantages, for instance, the simple and unambiguous interpretation of the IC diagram that is important for comparison with the empirical data. As shown above, the results of comparison not only approve the new concept, but can also be used to evaluate the relative significance of various factors (besides the orbital element variations) controlling the climate of the planet. The new data obtained may be applied as well in the refined chronology of climatic events.

We see the further development of the concept in searching for concrete mechanisms transforming the orbit-dependent insolation variations into climatic changes and in obtaining their mathematical formulation. Apparently, this will be a difficult task in the case of insolation variations dependent on precession and obliquity.

One more important direction in the development of the new concept is the correlation of the IC diagram with a wide spectrum of paleoclimatic data, including those obtained from studies on the continental deposits. Investigation of continental records is necessary for the more comprehensive and reliable understanding of global climatic changes. This general scientific problem cannot be solved without joint efforts of different scientists.

Principal conclusions of the study are as follows.

(1) Numerical data on insolation variations during caloric semiannual periods within different latitudes, as they have been calculated by Milankovitch, cannot be used for the interpretation of real climatic changes, because the spatial and temporal relationships have been underestimated in his calculation and interpretation of insolation magnitudes.

(2) Small in amplitude but long in duration insolation changes connected with the eccentricity variations had an independent climatic influence during the Pleistocene.

(3) The newly plotted IC diagram is well correlative with the OI records. This result represents first direct evidence demonstrating the decisive influence of the orbital elements variations on the Pleistocene climate.

(4) The general trend and rhythms of global climatic changes during the last million years have been mostly controlled by the summary insolation changes dependent on variations of orbital elements, among which the eccentricity contribution was of prime significance. The considerable climatic effect of orbital element variations and some peculiar features of climatic changes during the time period in question turned out to be realized because of certain factors related to the evolution of the planet and its climatic system.

(5) The above conclusions essentially change the former concept of the orbital elements impact on climatic changes and may be considered as basic postulates of the new astronomical theory of the paleoclimate.

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The Jatulian (Lower Proterozoic) Carbonates with Abnormally Heavy Carbon: Stratigraphic Position and Paleogeographic Distribution in the Baltic Shield

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Abstract—Vertical and lateral variations of $\delta^{13}C$ and $\delta^{18}O$ in the Lower Proterozoic (2.3–2.0 Ga) carbonates of the Baltic Shield are analyzed on the basis of carbon and oxygen isotope measurements in 460 Jatulian rock samples. The most heavy carbon isotope composition ($\delta^{13}C > 10‰$) is recognized in the Middle and basal Upper Jatulian parts of carbonate sequences (lower beds of the Upper Jatulian according to the two-member subdivision scheme). Accumulation age limits for carbonates with abnormally heavy C-isotope composition are estimated to be 2.25 and 2.08 Ga; the peak values of this isotopic parameter are confined to 2.12–2.10 Ga. The maximum average values estimated for lateral distribution of $\delta^{13}C_{carb}$ are concentrated in deposits of the axial zone of the Jatulian basin and decrease toward its periphery down to $\delta^{13}C < 5‰$. Any section with carbonates yielding the abnormally heavy C-isotope composition appears to be correlative with the Jatulian interval. The opposite statement, that a section where carbonates of this kind are missing cannot be correlated with this interval, is incorrect. The established peculiar geographic distribution of $\delta^{13}C$ and $\delta^{18}O$ in carbonates of the Baltic shield suggests that similar distribution patterns may be characteristic of other regions with Precambrian rocks. These distribution patterns were apparently controlled by geodynamic and paleogeographic factors in the course of the planet evolution.

Key words: *Lower Proterozoic, Baltic shield, carbonates, heavy carbon, distribution, basinal control*

After the discovery of heavy carbon isotope composition in the Lower Proterozoic carbonates of the Lomagundi province of South Africa (Schidlovski *et al.*, 1976), similar rocks have been recognized at the same stratigraphic level of Precambrian sections in four continents (Melezhik *et al.*, 1997), and this points to the global significance of the carbon isotope anomaly in deposits of about 2.3–2.0 Ga old. The above conclusion supported by many researchers (e.g., Yudovich *et al.*, 1990; Tikhomirova and Makarikhin, 1993; Melezhik and Fallick, 1994; Akhmedov, 1996; Akhmedov *et al.*, 1996) can be of a high stratigraphic potential for the Lower Proterozoic, where we operate so far with a limited series of regional, interregional, and global stratigraphic markers.

The Baltic shield is one of the best studied Precambrian regions. Its eastern part, up to 800000 km^2 in area, experienced the Archean cratonization and hosts many Early Proterozoic synclines and synclinoria with carbonates of the Jatulian or close age, which yield $\delta^{13}C$ parameters approaching 10–18‰ (Yudovich *et al.*, 1990; Karhu and Melezhik, 1992; Karhu, 1993; Akhmedov and Krupenik, 1995; Melezhik and Fallick, 1996). The data base of more than a thousand of C-isotope analyses for rocks from various lower Precambrian stratigraphic levels of the Baltic shield shows that almost half of the results characterizes carbonates displaying an abnormally heavy carbon isotope composition. This situation is favorable for understanding in detail some stratigraphic and paleogeographic aspects of these intriguing rocks in the region under consideration.

Sections studied in detail within Karelia (Russia), for instance, in the Onega synclinorium, include rocks with a abnormally heavy C-isotope composition at all stratigraphic levels of the Jatulian (Yudovich *et al.*, 1990; Tikhomirova and Makarikhin, 1993; Akhmedov and Krupenik, 1995; Melezhik *et al.*, 1996), although the highest values of $\delta^{13}C$ are established in carbonates of the basal Upper Jatulian. According to limited data on the Lower Jatulian rocks (Tikhomirova and Makarikhin, 1993), $\delta^{13}C$ values in this interval are close to +5‰. In deposits underlying or overlying the Jatulian, a heavy carbon isotope composition was established in a few cases only.

The publications listed above reported, in addition, that carbonates with a comparable carbon isotope composition occur outside Karelia within the rock sequences well correlated to the Jatulian on the basis of geological data (Table 1). Moreover, in the Kiihtelysvaara and Kuusamo areas of Finland, from where the Jatulian structures extend eastward into Karelia, the

Table 1. Stratigraphic position of carbonates with the heavy C-isotope composition in some sections of the Baltic shield

	Geological structures						
	Onega	Tulomozero	Kiihtelysvaara (Pekkarinen and Lukkarinen, 1991)	Kuusamo (Silvennoinen, 1972, 1991)	Perapohja (Perttunen, 1985, 1991)	Pechenga	Imandra-Varzuga
	Ludicovian		Petaiko	Phyllite	Martimo	Pilgujarvi	
Jatulian	Upper (**Tulomozero**)	*Upper Tulomozero*	*Annala*	*Limestone-dolomite* Amphibole schist	*Rantamaa*	Kolasioki	Il'mozero
		Lower Tulomozero	***Viistola***	***Dolomite*** ***Quartzite***	*Tikanmaa*	***Kuetsjarvi***	***Umba***
			Kalkunmjaki		***Kvartsimaa***		
	Middle (Medvezh'egorsk)		Koljola	Greenstone III ***Siltstone*** Greenstone II	Eoutiaapa ***Kivalo*** Runkaus		Polisar
	Lower (Jangozero)		Haukilampi Viesimo	*Sericite schist*	*Sompujarvi*		Seid-Rechka
	Sariolian			Conglomerate		Ahmalahta	Kuksha
	Sumian						
	Archean						

Note: Table columns correspond to local sections of subdivisions (formations of Finland) indicated in the top row; carbonate levels with high anomalous and anomalous $\delta^{13}C$ values are italicized in bold and normal respectively.

high $\delta^{13}C$ values are characteristic, according to Karhu (1993), not only of the Upper, but also of the Middle Jatulian carbonates (the case of three-member subdivision scheme) missing from the Karelian sections. The same event is recorded in the Perapohja structure of Finland. At the same time, some sections of the condensed facies type produce a perception that rocks with the heavy carbon isotope composition occupy other stratigraphic levels. For instance, within the Tulomozero structure of Karelia, the highest $\delta^{13}C$ values are recorded in the basal portion of the Jatulian interval. The interval begins here with the Tulomozero Formation of the Upper Jatulian because of the pinch-out of Lower–Middle Jatulian sedimentary facies, and, disregarding this fact, one may erroneously interpret the stratigraphic position of the rocks in question as corresponding to the Lower Jatulian.

A statement based on fallacy is exemplified in works considering the stratigraphic position of carbonates with the anomalous carbon isotope composition in the Pechenga–Varzuga belt of the Kola Peninsula (Karhu and Melezhik, 1992; Pokrovskii and Melezhik, 1995, Melezhik and Fallick, 1996), where the high $\delta^{13}C$ values are mostly concentrated in sediments of the Kuetsjarvi and Umba formations correlated in the cited publications with the Lower Jatulian of Karelia. Such a correlation is not unambiguous, however. In the Pechenga structure, the Ahmalahta Formation is thought to be an equivalent of the Sariolian of Karelia, whereas the Kuetsjarvi and Kolasioki formations are regarded as analogs of the Lower and Upper Jatulian, respectively. The isotopic anomalies are established mostly in the lower and, to a lesser extent, in the upper parts of the Kuetsjarvi Formation, which are represented by three rhythmical members. The detailed characteristics of the members and their stromatolites (*Magmatism...*, 1995) suggest that they correspond to the Upper Jatulian deposits of Karelia. The lower part of the Kolasioki Formation is also correlative with the Upper Jatulian, whereas its upper part enclosing the carbonaceous beds is of the Ludicovian age. Consequently, geological, lithological, and isotopic data on the Pechenga segment of the belt do not contradict the general idea on the stratigraphic position of the anomalous carbonates characterized above.

There are two stratigraphic schemes accepted for the Imandra–Varzuga structure. In one of them, the Seid-Rechka Formation is attributed to the Sumian, because the polymictic conglomerates of the overlying Polisar Formation are thought to be identical to those of the Sariolian (*Imandra-Varzugskaya...*, 1982), and also because an "imandtrite," the acidic granophyre of the definite contact origin, yielded the isotopic dates exceeding 2.4 Ga (Melezhik *et al.*, 1997) and was considered as a rock of the Seid-Rechka Formation. In such a scheme, the Umba Formation overlying the Polisar Formation and enclosing carbonates with the heavy carbon isotope composition is "automatically" placed at the Lower Jatulian level.

Another scheme is based on the evident similarity between sections of the Kuksha and Ahmalahta formations of Pechenga area, which are correlated to each other and to the Sariolian of Karelia (Golubev *et al.*, 1984; Heiskanen, 1990). In this version, the quartzite and volcanic beds of the Seid-Rechka Formation are attributed, together with the basal weathering crust, to the Lower Jatulian, while the polymictic and volcanogenic rocks of the Polisar Formation are correlated with its middle subdivision. The Umba Formation and basal beds of the Il'mozero Formation are considered as analogues of the Upper Jatulian Tulomozero Formation. This version is better substantiated in its geological aspect and does not contradict the general point that the carbon isotope anomaly is confined to the Upper Jatulian of the entire region, the Pechenga–Varzuga belt included.

Thus, the revised stratigraphic position of carbonates marked by a heavy carbon isotope composition shows their universal confinement to the Middle–Upper Jatulian interval (or to the upper one in the two-member subdivision scheme) of the Lower Proterozoic of the Kola Peninsula and Karelia, if the lithology of deposits, principal geological markers, bed successions, and key relations between sedimentary and magmatic complexes are properly taken into account (Heiskanen, 1990, 1992). The same conclusion appears to be correct with respect to deposits of this level outside of the region under consideration.

The lateral distribution of carbonates yielding the anomalous carbon isotope composition is analyzed on the basis of 460 results (Table 2) of C- and O-isotope measurements (Yudovich *et al.*, 1990; Tikhomirova and Makarikhin, 1993; Karhu, 1993; Akhmedov and Krupenik, 1995; Melezhik and Fallick, 1996; Melezhik *et al.*, 1996). Many analytical data on high and moderate $\delta^{13}C$ values are obtained for the densely sampled drill-hole sections, thus excluding a systematic error in processing the particular data samples that would be possible in the case of selective exposure degree of the studied objects. This is also evident from the relative stability of the dispersion parameters calculated for all data samples and lacking any correlation with the size of the latter. The selective distribution characteristics of $\delta^{13}C$ and $\delta^{18}O$ values in carbonates show them to be fitting the normal distribution law and thus substantiate the applicability of the small-sample method for the calculation of average values and their subsequent analysis. The average $\delta^{13}C$ value calculated for all 460 analytical results is +7.98‰, and the anomalous carbon isotope composition appears to be typical of the studied region as a whole. The data subjected to the processing characterize the Upper Jatulian carbonate deposits (Tulomozero Formation) of the Karelian Complex recorded in the regional stratigraphic scheme of the Kola–Karelian region, and also the correlative rocks known in Finland, Sweden, and Norway (Melezhik and Fallick, 1996; Heiskanen, 1990, 1992).

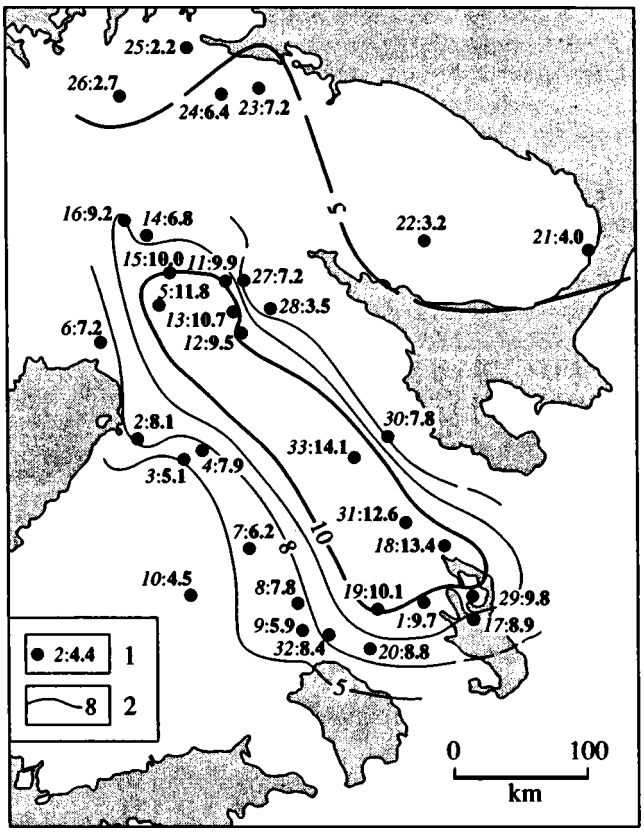

Fig. 1. Scheme of $\delta^{13}C$ isolines for the Jatulian (Lower Proterozoic) carbonates of the Baltic shield: (1) number of sampling site and average $\delta^{13}C$ (in Fig. 1) and $\delta^{18}O$ (in Fig. 2) value; (2) isolines of average $\delta^{13}C$ values.

The original data have been grouped first into samples characterizing 33 particular areas, and average values of isotopic parameters in carbonates were calculated for each area. Afterward, these values were used to plot isolines in topographic schemes (Figs. 1, 2). The schemes obtained are rather simple: they both show the extended zones of the northwestern orientation, which delineate the high average values and are surrounded by zones of the relatively low, though still anomalous parameters ($\delta^{13}C > +5.0‰$) of carbon isotope composition. The statistical analysis of data samples shows that the random average $\delta^{13}C$ value within the axial zone is greater than the general average yielding a probability coefficient greater than 0.99, whereas the average $\delta^{13}C$ for point 23 is significantly lower (a probability coefficient more than 0.95) than the general one. Thus, the obtained scheme of isolines depicts contours that are meaningful in terms of statistics and separating areas of high and low $\delta^{13}C$ values in the analyzed data samples. The same conclusion is correct with respect to the scheme of $\delta^{18}O$ isolines.

Because of some data deficiency for central areas of both schemes, isolines around one (Fig. 1) or two separate (Fig. 2) axial zones of high isotopic parameters appear to be conditional to a certain extent, but this

Table 2. Isotopic parameters of Upper Jatulian carbonates and their analogues from the Baltic shield

No.*	n**	$\delta^{13}C$***		$\delta^{18}O$****		Sampling site	Formation	Reference
		X_c	Y_c	X_0	Y_0			
1	59	9.66	5.66	21.44	4.86	Pjalozero	Tulomozero	1(39), 2(7), 3(5), 5(8)
2	1	8.12		17.23		Muhos		3(1)
3	6	5.11	4.82	19.15	4.73	W. Kainuu	Horkankallio, Eskosenvaara	3(5)
4	4	7.9	0.11	17.41	2.14	E. Kainuu	Hyrynsalmi	3(4)
5	9	11.76	2.39	18.8	6.11	Misi		3(9)
6	21	7.23	7.46	20.18	5.79	Perapohja	Kivalo, Kvartsimaa, Rantamaa	3(21)
7	5	6.19	3.77	16.63	4.51	Juuka	Viistola	3(5)
8	10	7.84	6.17	19.04	3.38	Kiihtelysvaara	Viistola	3(10)
9	5	5.89	0.35	16.52	5	Sortavala	Lower Pitkjaranta	3(2)
10	13	4.55	3.67	17.94	4.78	Kuopio	Petonen	3(12)
11	3	9.9	7.88	19.7	3	Salla	Aatsinginhauta, Matovaara, Kelloselka	3(3)
12	3	9.5	5.08	14.22	0.45	S. Kuusamo	Saricite schist, Siltstone	3(3)
13	7	10.66	5.35	19.45	2.78	N. Kuusamo	Dolomite	3(10)
14	6	6.8	4.81	13.06	4.51	Sodankyla	Middle Lapponian	3(6)
15	5	10.03	7.41	16.55	4.82	Pelkosenniemi	Middle Lapponian	3(5)
16	7	9.23	8.42	16.59	3.21	Kittila	Upper Lapponian	3(7)
17	4	8.88	0.81	24.2	6.32	Olenii Is.	Tulomozero	3(2), 5(2)
18	4	13.43	4.86	17.57	8.55	Medvezh'egorsk	Tulomozero	3(4)
19	2	10.06		20.06		Suojarvi	Tulomozero	3(2)
20	6	8.85	4.22	18.49	2.94	Tulomozero	Tulomozero	3(6)
21	9	4.04	2.11	16.59	5.71	Ust-Ponoi	Carbonate 2 and 3	4(4)
22	11	3.19	7.1	13.17	3.74	Imandra–Varzuga	Umba	4(11)
23	120	7.25	1.34	17.03	6.98	Pechenga	Kuetsjarvi	4(120)
24	10	6.41	2.21	16.19	15.52	Pasvik	Koievann	4(10)
25	3	2.24	0.96	12.98	0.09	Polmak	Vuopp'ejavri	4(3)
26	2	2.7		16.17		Karasjok	Gallebai'ke	4(2)
27	26	7.18	7.76	16.16	6.54	Kuolajarvi	Sovajarvi	4(26)
28	11	3.5	1.23	21.42	0.92	Sovajarvi	Sovajarvi	4(11)
29	72	9.81	5.33	22.3	2.41	Zaonezh'e	Tulomozero	6(72)
30	6	7.82	14.67	16.77	23.87	Lehta	Upper Jatulian	5(6)
31	5	12.57	0.65	22	5.38	Segozero	Upper Jatulian	5(5)
32	1	8.37		21.31		Janisjarvi	Tulomozero	3(1)
33	4	14.05	7.56	23.3	5.62	Chirka-Kem'	Tulomozero	1(4)

Notes: * Numbers as in figures;
** number of samples;
*** $\delta^{13}C$, ‰ PDB;
**** $\delta^{18}O$, ‰ SMOW; (X) average value; (Y) standard deviation.
References: 1, Akhmedov and Krupenik, 1995; 2, Yudovich *et al.*, 1990; 3, Karhu, 1993; 4, Melezhik and Fallick, 1996; 5, Tikhomirova and Makarikhin, 1993; 6, Melezhik *et al.*, 1996 (number of analyses is given in brackets).

does not change the essence of the results obtained. The first scheme seems to be especially interesting, because in the Lafoten Islands located at the northwestern continuation of the axial high-$\delta^{13}C$ zone, there are established the Lower Proterozoic carbonate deposits, which experienced a granulite-grade metamorphism. These rocks show an abnormally heavy carbon isotope composition as well (Baker and Fallick, 1989), although they have not been included in our sample for statistical analysis.

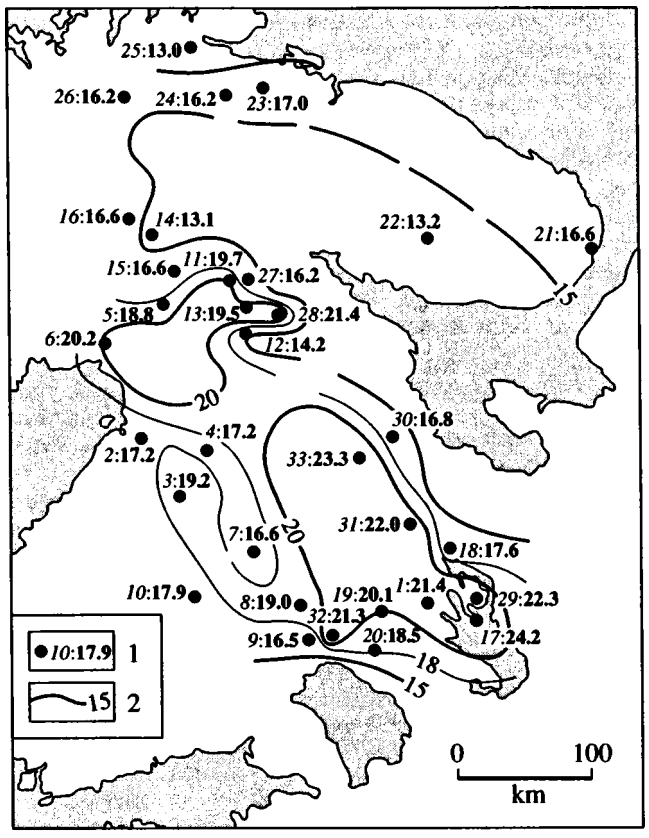

Fig. 2. Scheme of $\delta^{18}O$ isolines for the Jatulian (Lower Proterozoic) carbonates of the Baltic shield (symbols as in Fig. 1).

The comprehensive geological interpretation of the established distribution patterns of C- and O-isotope parameters in the Jatulian carbonates needs in additional data of further investigations. However even now, one can see that the established distribution patterns are independent of the configuration of the present Jatulian structures and Svecofennian metamorphic front. This situation and confinement of the anomaly maximums to the axial part of the shield, where cratonization occurred at the Archean time, suggest that the distribution patterns of the isotopic characteristics were controlled by the basin parameters of the Tulomozero time. Our inference is based on the assumption that the Late Jatulian sedimentary basin, being greater in size, all the same followed the general contours of the former Early Jatulian basin, which have been reconstructed on the basis of detailed lithological data (Heiskanen, 1990). If the large Late Jatulian basin was of evaporite type, as is stated in all publications cited above, then, from the theoretical standpoint, we should expect the $\delta^{13}C$ and $\delta^{18}O$ distribution patterns exactly like those established in this work.

It is a common opinion that the first shallow-water carbonate platforms have appeared in many Precambrian regions at the Jatulian time (2.3–2.1 Ga ago). Consequently, we may consider the mass accumulation of carbonates and their carbon isotope composition as a result of an obligate course of geodynamic development of the planet. There does not seem to be possible that the intense accumulation of carbonates enriched in the heavy carbon isotope is synchronous to the appearance of free oxygen in the atmosphere. The event that stimulates researchers to search for the fossilized organic matter may find an explanation in the unique biospheric processes that occurred in response to the climatic warming after the Sariolian Glaciation and to the related quick expansion of oceans and seas over the globe. Being interpreted in terms of the tremendous increment of the global biomass, the idea of "immortal" biota suggested by Yudovich *et al.* (1990, p. 976) is not as fantastic as it would seem at the first glance.

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Carbon Isotopes in Riphean Carbonates from the Yenisei Ridge

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Abstract—Carbon isotope composition is determined in the Riphean (1450–1350 to 850–820 Ma) carbonate rocks from the Sukhoi Pit, Tungusik, and Oslyanka groups of the Yenisei Ridge eastern zones. The results of petrographic, geochemical, and isotopic studies show that limestone and dolomite are mainly devoid of signs of significant postsedimentation alteration and could preserve the C-isotope signature close to the initial value. Carbonate rocks from the basal (Korda Formation) and upper (Sosnovyi Formation) parts of the Sukhoi Pit Group are characterized by the C-isotope ratios close to zero. Overlying deposits younger than 1000 Ma span the long intervals of their sequence, where $\delta^{13}C$ values are 2.5–4.5‰; with the extreems 5.6‰; these are separated by short intervals with lower $\delta^{13}C$ values. The Dzhur rocks show the ascending trend of $\delta^{13}C$ variations from 1.6–1.9 to 4.8‰ with the subsequent lowering of this parameter at the top of the formation. The Shuntarskaya Formation is characterized by positive $\delta^{13}C$ values (1.9–3.2‰) in its lower part and by negative values (–2.8–5.0‰) in the upper part. In the Angara section of the Seryi Klyuch Formation, values of 3.0 to 4.4‰ prevail, lowering sometimes to 1.2–2.6‰. Carbonate subformations of the Dadykta (Kamenka zone) and Rybnaya (Gorbilok zone) formations yield uniform $\delta^{13}C$ values (3.0–4.5‰) with their probable decrease at the top of the sections. The lower part of the Dashka Formation is marked by the ascending trend of $\delta^{13}C$ variations from 0.3 to 5.0‰; higher, there are variations from 2.3–3.0 to 4.5–5.6‰. The comparison of the $\delta^{13}C$ values characteristic of rocks from the Yenisei Ridge with those obtained for rocks from the Turukhansk uplift and Baikit arch shows that they are close in the first case and substantially differ in the second. $\delta^{13}C$ values in carbonate rocks of the Yenisei Ridge (except the Korda rocks) generally coincide with those recorded in 1100–850-Ma-old carbonates from other regions. Nonetheless, rocks from the Dadykta and Dashka formations with high (4–5‰) positive $\delta^{13}C$ values might be younger than 850 Ma.

Key words: *carbon isotopes, carbonate rocks, Riphean, Yenisei Ridge, stratigraphy, sedimentation environments*

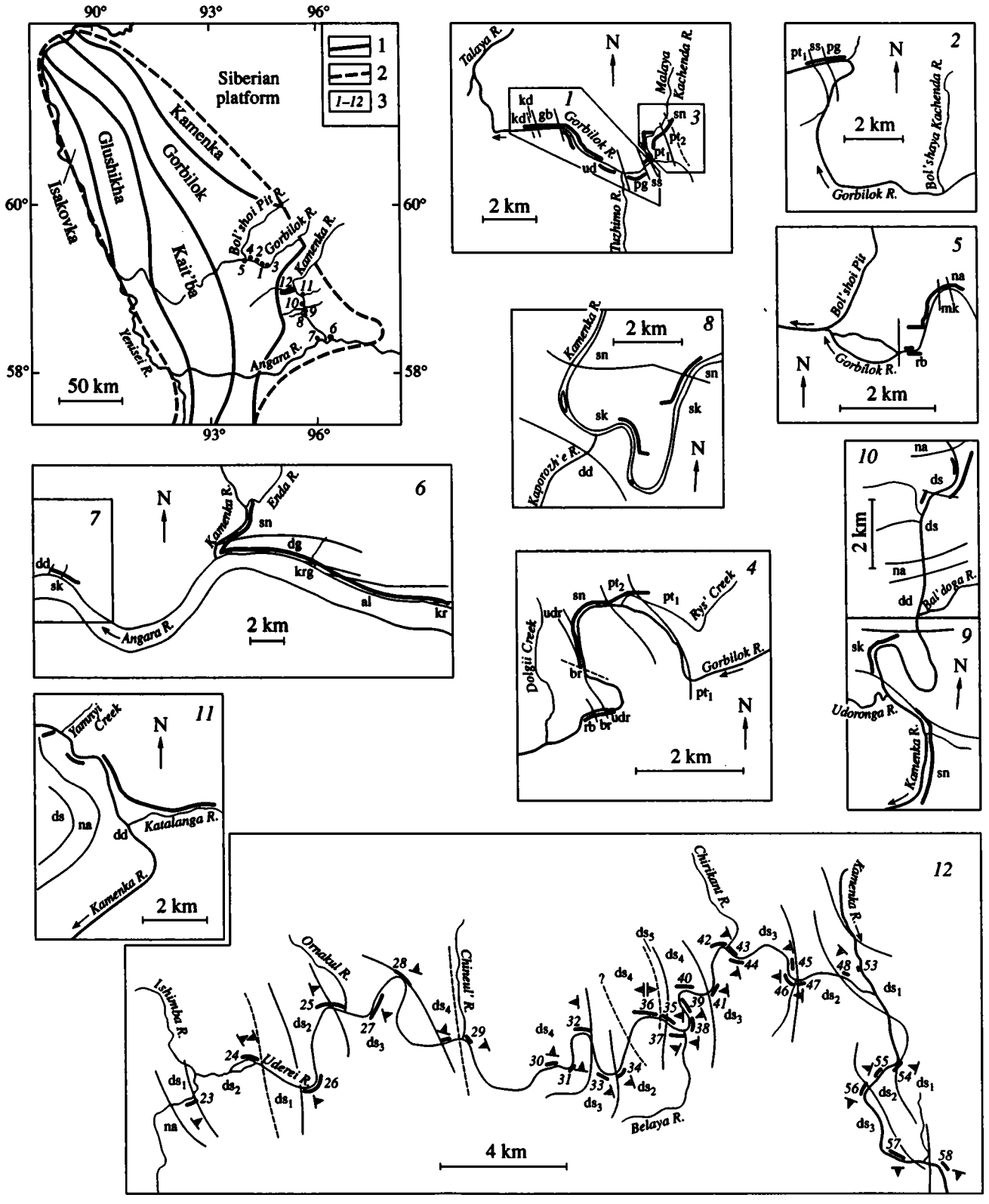
INTRODUCTION

Recently, abundant data have been obtained on the primary carbon isotope composition in marine carbonates. It is established that the geological record is characterized by variations in $\delta^{13}C$ values, which were controlled by the changing balance of organic and carbonate carbon in seawater. These variations are used to define the different-scale geological events and to solve a series of stratigraphic problems. Particularly important are chemostratigraphic studies on the Precambrian rocks, because traditional methods of subdivision and correlation are of a limited use in this interval.

The most significant results in the study on the carbon isotope behavior in the Precambrian carbonate rocks are obtained for the time interval of 850–550 Ma (Schidlowski *et al.*, 1975; Knoll *et al.*, 1986; Derry *et al.*, 1992; Pokrovskii and Gertsev, 1993; Pokrovskii, 1996; Ripperdan, 1994; Strauss *et al.*, 1994; and others). They indicate that this period was characterized by high $\delta^{13}C$ values (to 4–8‰), which sometimes fall to –2–0‰. Less known is the distribution of carbon isotopes in carbonate rocks from older Riphean strata (Schidlowski *et al.*, 1975; Pokrovskii and Vinogradov, 1991; Ghazban *et al.*, 1992; Veizer *et al.*, 1992; Knoll *et al.*, 1995; Kah and Knoll, 1996; Khabarov *et al.*, 1996; Podkovyrov and Vinogradov, 1996). According to these data, the interval of 1600–1200 Ma is marked in general by $\delta^{13}C$ values close to zero, whereas younger deposits reveal variations in the C-isotope composition through the section with gradual enrichment in ^{13}C. Despite the progress in studies, the detailed $\delta^{13}C$ record is not yet established for this period of time primarily because of inadequate isotopic data on the particular Riphean sequences. In this relation, the thick, mainly carbonate section in the eastern zones of the Yenisei Ridge, which is lithologically well studied (Khabarov and Tanygin, 1993; Khabarov, 1994; Khabarov *et al.*, 1998), might be well informative. The results of lithological, petrographic, and isotopic studies on carbonate rocks from the Yenisei Ridge are discussed below. They provide grounds for defining the evolutionary trend of the carbon isotope variations and a possibility for applying this trend for stratigraphic correlation of Riphean sequences.

Regional Geological Structure and Stratigraphy

The Yenisei Ridge represents a complex fold and thrust structure formed, mainly, during the Early Baikalian collision (800–850 Ma ago) and complicated in the course of later geological events (Semikhatov, 1962; Votakh, 1968; Postel'nikov, 1980; Khabarov, 1995; Kontorovich *et al.*, 1996; Surkov *et al.*, 1996; Khomentovskii, 1996).



The Riphean (pre-Baikalian) structural complex comprises five structural–formational zones of a submeridional strike: the Kamenka, Gorbilok, Kait'ba, Glushikha, and Isakovka zones (Fig. 1). The Riphean is represented by the thick (10–12 km) laterally variable complex of siliciclastic, carbonate, and volcanogenic–sedimentary deposits. The latter are mainly confined to western zones. The section consists of four units (form the base upward): the Teya, Sukhoi Pit, Tungusik, and Oslyanka groups, which are overlain with the angular unconformity by the Upper Proterozoic (Baikalian and Vendian) deposits (Fig. 2). The Riphean age of the Teya Group is still debatable (Votakh, 1968; Kovrigina *et al.*, 1975; Volobuev *et al.*, 1976; Postel'nikov, 1980; Shenfil', 1991). According to our data (Khabarov, 1994), the upper part of the Penchenga terrigenous–carbonate formation of the Teya Group in the Gorbilok zone (Chirimba River) is connected to the overlying Korda Formation of the Sukhoi Pit Group and forms a single sedimentary complex together with the true Riphean deposits.

The Sukhoi Pit Group comprises six lithostratigraphic units: the carbonate–terrigenous Korda Formation (to 700 m), the green fine-grained siliciclastic Gorbilok Formation (750–900 m), the fine-grained siliciclastic Uderei Formation with a member of clayey limestone in the middle part (to 1200 m), the sandy–silty–clayey Pogoryui Formation (to 750 m), the clayey–limy Kartochki Formation (to 150 m), and the dolomite Alad'ino Formation (to 500 m). In the Gorbilok zone, the two latter formations are replaced by the clayey–carbonate Sosnovyi Formation to 350 m thick (Semikhatov, 1962).

The Tungusik Group conformably or, according to some data (e.g., Shenfil', 1991), unconformably rests upon the Sukhoi Pit Group and is composed in the Kamenka zone of five lithostratigraphic units: the siliciclastic Krasnaya Gora (60–140 m), carbonate Dzhur (to 400 m), the dark-colored terrigenous–carbonate Shuntarskaya (to 1000 m), the carbonate Seryi Klyuch (to 600 m), and the terrigenous–carbonate Dadykta (to 600 m) formations. In the Gorbilok zone, the Tungusik Group is subdivided into six units (Fig. 2): the carbonate–terrigenous Potoskui (to 1050 m), dark-colored carbonate–clayey Shuntarskaya (to 500–600 m), siliciclastic Udoronga (to 120 m), carbonate–clayey Borema (to 150 m), carbonate–terrigenous Rybnaya (to 480 m), and fine-grained siliciclastic Mokryi (to 270 m) formations.

The basal unit of the Oslyanka Group corresponds to the Nizhneangarsk Formation (to 350 m), which rests on older rocks conformably or with local erosion marks and is overlain by the limy Dashka Formation up to 1500 m thick.

The lithological study of Riphean deposits was carried out in many sequences exposed along the banks of the Angara, Kamenka, Gorbilok, Bol'shoi Pit rivers and their tributaries. In some areas of the Gorbilok zone, with the complex geological structure, the study was accompanied by geological mapping.

Samples for isotopic analysis were collected only from the most representative sequences (Fig. 1). The sampling step in continuous carbonate sections averaged 20–25 m. Unfortunately, some formations were sampled irregularly. Sometimes, for instance, in closely spaced sections of carbonate rocks with different degrees of recrystallization and dolomitization, samples were taken with a step of 2–5 m, whereas in other cases, the step was 50 m. The dolomitic Korda Formation was sampled from the exposure located on the right bank of the Gorbilok River, 2.5 km upstream from the Talaya River mouth. Further to the Malaya Kachenda River mouth, the exposed higher strata correspond to the Sukhoi Pit Group (Figs. 1.1, 2.1). Nevertheless, limestone samples of the Sosnovyi Formation were collected from the more representative section exposed on the right bank of the Gorbilok River, 7.5–8.0 km downstream from the Bol'shaya Kachenda River mouth. The most complete section of the Potoskui Formation is exposed along the Gorbilok River, upstream from the Malaya Kachenda River mouth and near the latter, where limestone samples were taken from the upper Potoskui Subformation (Figs. 1.3, 2.3). The younger deposits of the Tungusik Group of the Gorbilok zone are relatively well exposed in the lower reaches of the Gorbilok River below the Rys' Creek mouth (Figs. 1.4, 2.4) and near the mouth of the Gorbilok River (Figs. 1.5, 2.5).

Samples of carbonate rocks for C- and O-isotope studies from the Dzhur Formation were taken from its stratotype section, located on the right bank of the Angara River, upstream from the Kamenka River mouth. In the mouth area of the latter, there is a transi-

Fig. 1. Scheme of the structural–formational zoning of Riphean deposits in the Yenisei Ridge and location of examined sections: (1) boundaries between zones; (2) the Yenisei Ridge boundary; (3) location of sections shown in inset maps. Sections: (*1*) Gorbilok River downstream from the Malaya Kachenda River mouth, (*2*) Gorbilok River, 7–8 km downstream from the Bol'shaya Kachenda River mouth, (*3*) Malaya Kachenda River 0–2 km upstream from the mouth, (*4*) lower reaches of the Gorbilok River downstream from the Rys' Creek mouth, (*5*) Gorbilok River, 2–4 km downstream from the mouth, (*6*) Angara River upstream from the Kamenka River mouth and the Kamenka River downstream from the Enda River mouth, (*7*) Angara River 16 km downstream from the Kamenka River mouth, (*8*) Kamenka River 1–6 km downstream from the Kaporozh'e River mouth, (*9*) Kamenka River near the Udoronga River mouth, (*10*) Kamenka River upstream from the Bal'doga River mouth, (*11*) mouth area of the Katalanga and Kamenka rivers, upstream, (*12*) lower reaches of the Uderei and Kamenka rivers, downstream. Formation indices in inset maps: (kd) Korda, (gb) Gorbilok, (ud) Uderei, (pg) Pogoryui, (ss) Sosnovyi, (kr) Kartochki, (al) Alad'ino, (pt) Potoskui, (krg) Krasnaya Gora, (dg) Dzhur, (sn) Shuntarskaya, (udr) Udoronga, (br) Borema, (sk) Seryi Klyuch, (rb) Rybnaya, (mk) Mokryi, (dd) Dadykta, (na) Nizhneangarsk, (ds) Dashka.

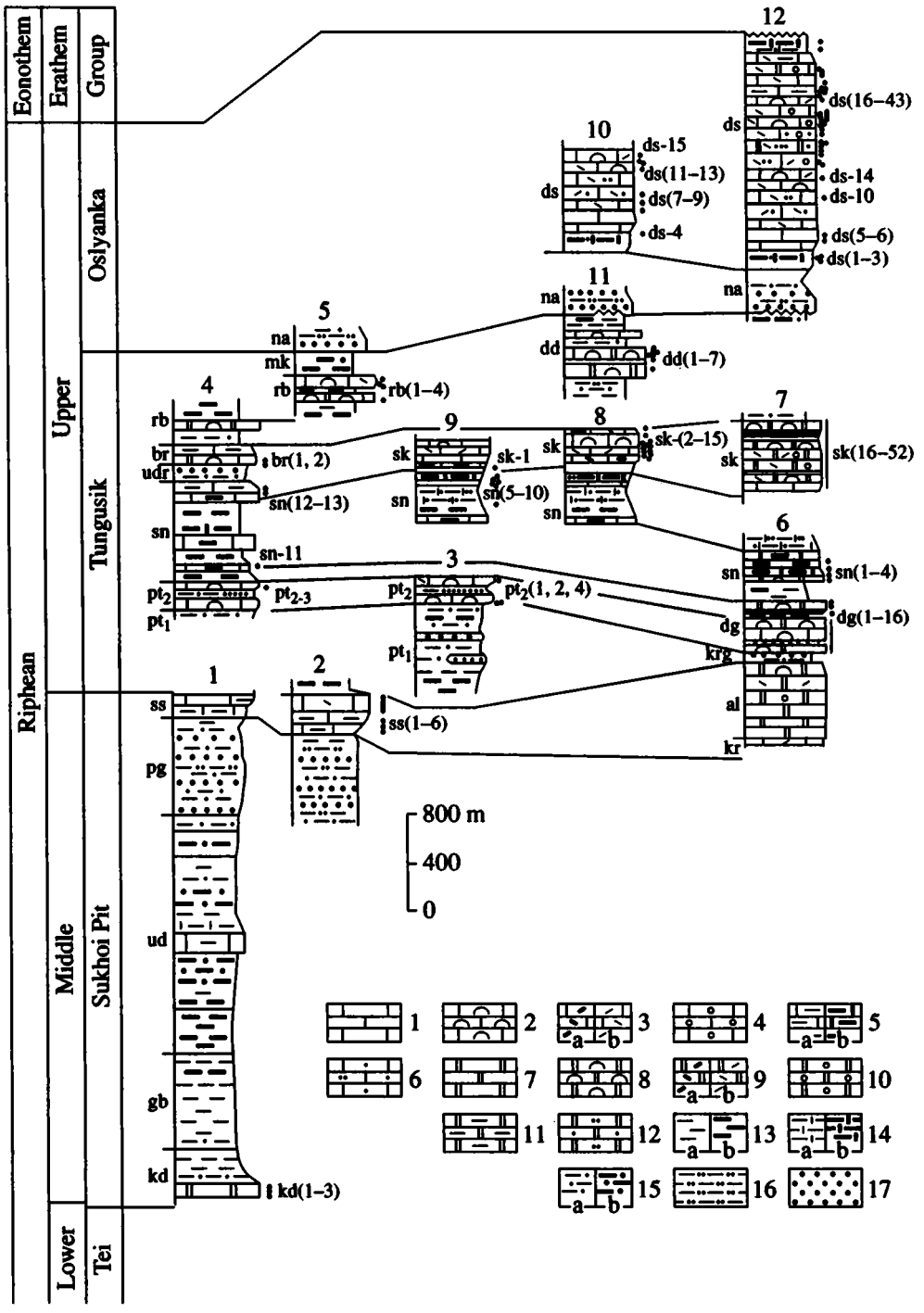


Fig. 2. Correlation of Riphean sections in eastern zones of the Yenisei Ridge and sampling sites for the C-isotope analysis. (1–6) limestones: (1) siltite–micrite, partly recrystallized, (2) stromatolitic, (3) coarse-intraclastic (a) and fine-intraclastic (b), (4) oolitic–lumpy–pisolitic, (5) clayey (a) and clayey–carbonaceous (b), (6) sandy and silty; (7–12) dolomites: (7) siltite–micrite, partly recrystallized, (8) stromatolite, (9) coarse-intraclastic (a) and fine-intraclastic (b), (10) oolitic–lumpy–pisolitic, (11) clayey, (12) sandy and silty; (13–15): argillites: (13) low-carbonaceous (a) and carbonaceous (b), (14) calcareous (a) and calcareous–carbonaceous (b), (15) silty (a) and silty carbonaceous (b); (16) siltstone; (17) sandstone. For numbers above columns and formation indices see Fig. 1.

tion to the overlying Shuntarskaya Formation, whose layers are traced to the Enda River mouth (Figs. 1.6, 2.6). The upper part of the Shuntarskaya Formation and overlying Seryi Klyuch Formation are exposed along the Kamenka River, 1–6 km downstream from the Kaporozh'e River mouth (Figs. 1.8, 2.8) and also around the Udoronga River mouth (Figs. 1.9, 2.9). The Shuntarskaya Formation was sampled in the last site

and the overlying Seryi Klyuch Formation, in the former. In addition, we thoroughly sampled the stratotype section of the Seryi Klyuch Formation, which is exposed on the right bank of the Angara River, 16 km downstream from the Kamenka River mouth (the Shuntarskaya stony river bed) (Figs. 1.7, 2.7). The upper, mainly carbonate part of the Dadykta Formation was studied in the mouth area of the Katalanga River and also along the Kamenka River, 2–4 km upstream from the Katalanga River mouth (Figs. 1.11, 2.11).

The most complete, though also fragmentary section of the Dadykta Formation is observed in the lower reaches of the Uderei River on the western and eastern limbs of the Nizhnii Uderei syncline. Additional information on the structure of the Dashka Formation was obtained from its sequences exposed along the Kamenka River, downstream from the Uderei River mouth (Figs. 1.12, 2.12) and 3–4 km upstream from the Bal'doga River mouth (Figs. 1.10, 2.10). Rocks for C- and O-isotope studies were sampled from several exposures (23, 25, 28, 30, 32, 35, 36, 40, 42, 43, 46, Fig. 1.12), which likely allowed us to perform sampling throughout the entire Dashka Formation (Figs. 1.12, 2.12). In addition, samples were also collected from the lower part of this formation, exposed along the Kamenka River upstream from the Bal'doga River mouth (Figs. 1.10, 2.10). The detailed characteristic of carbonate deposits exposed in the eastern zones of the Yenisei Ridge, where samples for isotopic studies were collected, is given below.

In the Korda Formation, carbonate rocks occur in its lower part, where they are represented by gray and greenish gray recrystallized, partly silicified dolomites with an admixture of siliciclastic material.

The Kartochki Formation (150 m) is composed of greenish gray and variegated thin-bedded micritic siltite and limy argillite, which form thin layers of distal turbidites and hemipelagic rocks; there are also rocks representing deposits of debris flows. Higher, the section consists of upper-slope and shelf thin-bedded dolomites with interbeds of synsedimentary slumps and postsedimentation breccia, oolitic silicified dolomites, and microbial laminites of the Alad'ino Formation (500 m). In the Gorbilok zone, these formations are replaced by the Sosnovyi Formation (to 350 m) composed of limestone intercalated with clayey–limy gravitites and hemipelagic deposits (Tanygin and Khabarov, 1989; Khabarov, 1994).

The Dzhur Formation (to 400 m) of the Kamenka zone is largely composed of limestones and dolomites of variable texture and genesis, which accumulated within the carbonate platform with marginal stromatolitic reefs. The basal part of the formation is represented by stromatolitic and intraclastic dolomites, which grade into a thin package of siliciclastic rocks. Northwestward, the thickness of siliciclastic rocks increases to 30 m in the lower reaches of the Kamenka River (near the Glubokii Creek mouth) and to 70 m in the Gorbilok zone. Upsection, there are stromatolitic and intraclastic limestones (to 90 m) dolomitized in the upper part. Higher, these rocks grade into intraclastic and stromatolitic dolomites (to 60–70 m) up-section are replaced by argillites (to 40 m). The section terminates with a member of mostly stromatolitic dolomites (to 50 m) intercolated with limestone interbeds at its base and top. In the northwestern direction, the thickness of the carbonate members decreases, and the upper member of the stromatolitic dolomite is replaced by dark–gray, often calcareous argillites with thin interbeds of dark–gray clayey limestone. In the Gorbilok zone, the lateral analogues of the upper member of stromatolitic dolomite is included into the Shuntarskaya Formation (Khabarov and Tanygin, 1993; Khabarov, 1994). The upper Potoskui Subformation consists mostly of limy and clayey–limy gravitites and hemipelagic rocks separated by the argillite member. Synsedimentary uplifts are marked by horizons of siltstones, quartz sandstones, and stromatolitic limestones (Khabarov and Tanygin, 1993) (Fig. 2).

The Shuntarskaya Formation (500–1000 m) is represented by carbonaceous, strongly pyritized limestones with the variable C_{org} content (to 6–8%), by clayey limestones, and less commonly, by siltstones and dolomites. In the Kamenka zone, the formation is composed of three units approximately equal in thickness. The lower unit is argillitic; the middle one consists of clayey–carbonate rocks with stromatolitic dolomites at the base (to 18 m), which rapidly wedge out northwestward, and the upper unit is composed of carbonate–silty–clayey rocks with abundant packages of distal turbidites. Slumps and thin debris-flow deposits are observed. In the Gorbilok zone, clayey–carbonate turbidites are virtually missing, and the formation thickness is reduced to 500–600 m. The decrease in thickness of the Shuntarskaya Formation is particularly noticeable in the Gorbilok zone, since it includes here lateral analogs of both the upper part of the Dzhur Formation and the lower portion of the Seryi Klyuch Formation (Fig. 2) (Khabarov, 1994; Khabarov *et al.*, 1998).

The Shuntarskaya deposits accumulated mainly in the moderately deep stratified water basin with the O_2 deficiency and H_2S contamination in the bottom layer. On the east, the basin was bounded by a carbonate shelf, which occupied areas located mainly east of the modern eastern boundary of the Yenisei Ridge, whereas on the west, there were two systems of volcanic islands (the Glushikha back-arc and Isakovka island-arc systems). The main volume of carbonate material derived from the carbonate platform was deposited in the Kamenka zone by turbidity flows and bottom currents. In the Gorbilok zone, the hemipelagic sedimentation of clayey and carbonate–clayey material was preponderant. Sometimes, the oxygen-minimum zone extended to the deep shelf, favoring a good preservation of organic matter in shelf sediments (Khabarov, 1994; Kontorovich *et al.*, 1996; Khabarov *et al.*, 1998).

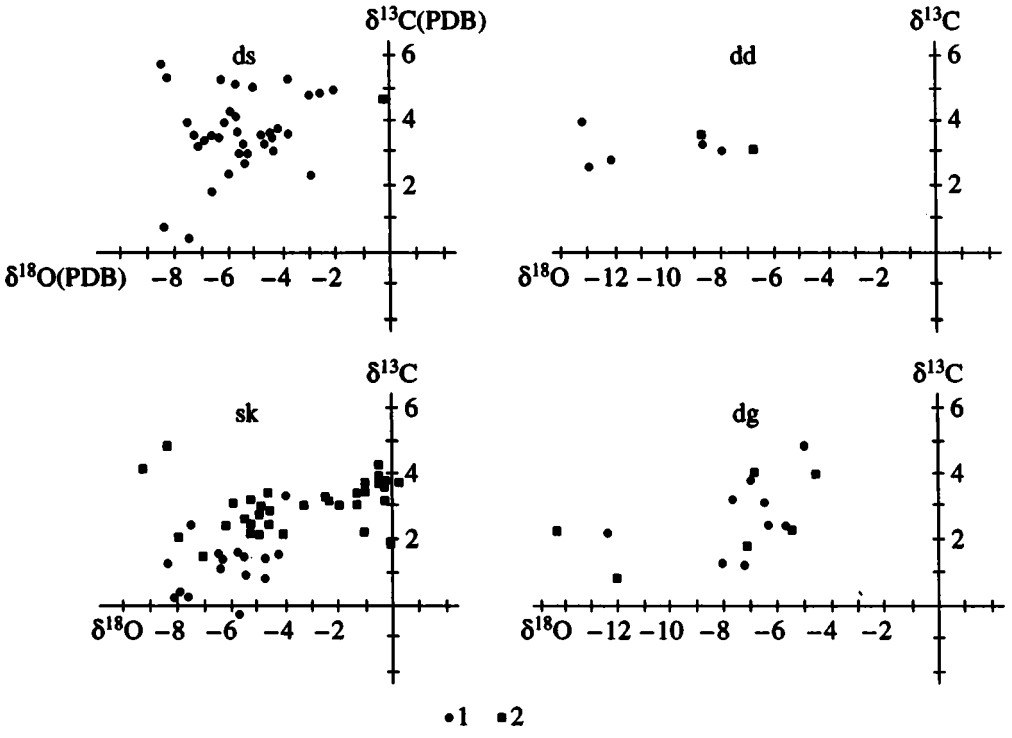


Fig. 3. Covariation of $\delta^{18}O$ and $\delta^{13}C$ in carbonate rocks from the Kamenka zone of the Yenisei Ridge: (1) limestone; (2) dolomite. For formation indices see Fig. 1.

The Seryi Klyuch Formation (600–400 m) and its lateral analogues are represented by three types of sections: dolomitic (Angara), limy (Kamenka) and limy–siliciclastic (Gorbilok). The first one is dominated by micritic–siltitic, oolitic–intraclastic, and stromatolitic dolomites deposited within the restricted shelf. Sections of the second type are composed of carbonaceous clayey micrites and micrite–siltitic limy turbidites in association with debris-flow deposits; upsection, they are replaced by subtidal and stromatolitic limestones, locally dolomitized. In the section of the third type, carbonate rocks (the upper part of the Shuntarskaya Formation and Borema Formation) occur together with argillites and silty sandstones of the Udoronga Formation to 120 m thick (Fig. 2), which were formed below and above the storm wave base. The interpretation of available data suggests that the Seryi Klyuch Formation was formed during the development of the carbonate platform with a marginal stromatolitic reef. Westward and northwestward, the carbonate platform graded into slope and basin, sometimes, anoxic settings; the latter experienced subsequent shoaling and accumulation of the siliciclastic material (Khabarov, 1994; Khabarov *et al.*, 1998).

The Dadykta Formation (to 600 m) in the Kamenka zone is mostly composed of stromatolitthic to oolitic–intraclastitic limestones and dolomites with the argillite horizon (to 280 m) in the lower part. Thin packages of siliciclastic rocks also occur within the largely carbonate part of the formation. In addition, the upper part of the formation encloses two argillite members separated by a stromatolitic limestone member (Fig. 2). In the Gorbilok zone, the role of argillite increases.

The Dashka Formation, up to 1500 m thick, consists mainly of limestones (micrite–siltite, clayey, oolitic–lumpy–intraclastic, stromatolitic) with siltstone, sandstone, and dolomite interbeds. Five units are distinguished in the formation. The basal unit (to 200 m) is represented by dark–gray limy and dolomitic argillites, with packages of gray to dark–gray micrite–siltite. The second unit (350–400 m) is composed of gray and greenish gray horizontally bedded and lenticular micrite–siltite, often silty and dolomitic; its upper part encloses oolitic–intraclastic and stromatolitic limestones. In easterly sections, the role of the latter rocks increases. The third unit (230–260 m) is dominated by gray micrite–siltite and oolitic–lumpy–intraclastic sandstones and dolomites with abundant syneresis fissures. Dolomites are usually sandy. Present are also siliciclastic sandstones. The fourth unit (410–440 m) is composed of gray and dark–gray siltite–micrite, oolitic–lumpy–intraclastic, and stromatolitic limestones, sometimes with an admixture of siliciclastic material and thin argillite members. The fifth unit (more than 160 m) is represented by dark–gray limy–dolomitic argillites with rare members of intraclastic limestone and silty–sandy dolomites. Rocks are usually grouped into frequent meter-thick cycles; they were formed in shelf environments under the distinct prevalence of storm processes.

Isotopic dates for Riphean deposits of the Yenisei Ridge are limited. Available are K–Ar glauconite dates for the Pogoryui Formation (about 1100 Ma), for the top layers of the Krasnaya Gora Formation (1007 ± 15 Ma), and for the base of the Dzhur Formation (924 ± 40 Ma) (Semikhatov and Serebryakov, 1983; Shenfil', 1991; and references therein). In addition, there are U–Pb dates of 850 and 850–820 Ma obtained for granitoids that intrude on the Tungusik Group deposits (Volobuev *et al.*, 1973).

Recently, the Pb–Pb date of 1035 ± 60 Ma was obtained for the Sukhaya Tunguska Formation of the Turukhansk uplift (Gorokhov *et al.*, 1995; Ovchinnikova *et al.*, 1995), which is correlative with the Alad'ino Formation (Semikhatov and Serebryakov, 1983; Shenfil', 1991; and others). These data indicate that carbonate rocks from the upper part of the Sukhoi Pit Group are not older than 1100 Ma and not younger than 1000 Ma, whereas the age of the Tungusik and Oslyanka groups is constrained by dates of 1000 and 850–820 Ma.

The age interpretation of carbonate rocks from the lower part of the Korda Formation and from the top of the Penchenga Formation is more difficult. Based on radiometric dating results, the age of the Teya Group metamorphism is estimated as 1850–1650 Ma (Volobuev *et al.*, 1976), but this date is inconsistent with geological observations in the Gorbilok zone (the Chirimba River, below the Chirimba granitoid massif). Deposits attributed in this zone to the Penchenga and Korda formations are interrelated by gradual transitions and show similar metamorphic alterations, which occurred during the formation of the Late Riphean granitoids. This inconsistency between geological and analytical data can be explained by the fact that the Penchenga Formation in different zones of the Yenisei Ridge includes heterochronous sequences, one of which is the Early Proterozoic and another is the Riphean in age.

Previously, we performed $^{40}Ar/^{39}Ar$ dating of synsedimentary diabases from the basal Riphean terrigenous sequence of the Baikit arch. These diabases are probably coeval with the Penchenga volcanism and intrusion of mafic rocks in the Rubnaya–Panimba structure of the Yenisei Ridge during the rifting (Khabarov, 1995). The data obtained indicate that the age of diabases is not less than 1430 ± 14(2σ) and not older than 1570 ± 27(2σ) Ma. Taking into consideration geological data, the base of the overlying Korda Formation is probably close to 1450–1350 Ma.

Analytical Procedure

Isotopic studies are performed after visual and microscopic examination of rocks in order to select the least recrystallized sample areas. Selected sample fragments were powdered and then treated by H_2O_2 for 20 minutes up to the decomposition of organic constituents. After drying, samples were decomposed using 100% othophosphoric acid under a temperature of 50°C. Gas fractions released during the first 2 and last 16 hours were sampled for measurements at the "Finnigan-delta" mass-spectrometer. The procedure allows one to separately determine $\delta^{18}O$ and $\delta^{13}C$ in readily soluble carbonate constituents of a single sample. The reproducibility of the OCO KH-2 (Germany) and MSA (Russia) standards did not exceed 0.1‰ for carbon and 0.15‰ for oxygen. To determine the isotopic composition of organic carbon, NBS-22 was used as a standard. The analytical results are presented relative to the PDB standard.

The Ca, Mg, Fe, and Mn contents in the soluble part of carbonate rocks were measured by the atomic absorption method using SP9 PI UNIKAM equipment (the error did not exceed 5%), and Sr concentrations are determined by the isotopic dilution method.

Postsedimentation Alteration

Many studies of carbonate rocks indicate that they can preserve the carbon isotope composition close to the initial one. Nevertheless, postsedimentation processes of different kind, can change the initial $\delta^{13}C$ values, usually lowering them (Veizer, 1983; Knoll *et al.*, 1995; Sochava *et al.*, 1996; and others). Therefore, the assessment of the postsedimentation rock alteration is considered to be a necessary step in stratigraphic isotopic–geochemical investigations, and pertinent methods are sufficiently well described (Brand and Veizer, 1981; Veizer, 1983; Derry *et al.*, 1992; Kaufman *et al.*, 1993; Kaufman and Knoll, 1995; Knoll *et al.*, 1995; Sochava *et al.*, 1996; and others). Samples were studied under the microscope to define areas with different petrographic characteristics. Mainly micrite, microsparite, oolitic, and stromatolitic rocks were selected for isotopic studies. However, this was sometimes impossible because carbonate rocks from eastern zones of the Yenisei Ridge usually show a variable degree of recrystallization. Several varieties are distinguished among the recrystallized carbonate rocks: (1) limestones and dolomites with weak micromaculose recrystallization and well-preserved grains; (2) dolomites (less commonly limestones) completely recrystallized, but with relics of the primary texture and sparites of later generation filling micropores and microcaverns; and (3) dolomites (less commonly limestones) represented by neomorphic crystals devoid of primary textural elements and oriented in parallel to foliation. According to petrographic–lithological data, dolomitization of limestones occurred mainly at the early diagenetic stage during intense exchange with seawater.

The interaction of carbonate rocks with meteoric and deep-seated waters is accompanied by negative $\delta^{18}O$ and $\delta^{13}C$ shifts; however, the water/rock ratio for carbon should be by 2–3 orders higher that for oxygen in order to produce similar changes in the isotopic composition. Therefore, the oxygen isotope ratios are considered as sensitive indicators of postsedimentation

Geochemical characteristic and carbon and oxygen isotope composition in Riphean carbonate rocks from eastern zones of the Yenisei Ridge

Sample	Zone	Level above the formation base, m	Mn, ppm	Fe, ppm	Sr, ppm	Mg/Ca	Mn/Sr	Fe/Sr	$\delta^{13}C_{carb}$ (‰ PDB)	$\delta^{18}O$ (‰ PDB)	$\delta^{13}C_{org}$
1	2	3	4	5	6	7	8	9	10	11	12
kd-1	Gorbilok	100	3500	44916.7	232.0	0.5021	15.0	192.7	–0.1	–10.9	
kd-2		120	2444.4	45250.0	214.4	0.4983	11.3	211.0	–0.8	–7.7	
kd-3		134	3812.6	22289.9	222.1	0.5064	17.7	164.0	–1.9	–10.4	
ss-1		60	1710.8	46927.7	540.0	0.0343	3.2	86.9	–0.5	–11.9	
ss-2		150	473.9	1825.8	399.1	0.0199	0.67	4.57	–0.4	–9.9	
ss-3		250	541.4	1473.8	482.8	0.0284	1.1	3.0	0.5	–11.0	
ss-4		270	319.2	1496.5	212.4	0.0032	1.5	7.0	0.7	–11.7	
ss-5		300	1089.4	10378.3	236.0	0.0422	4.6	43.9	0.7	–11.5	
ss-6		320	3700.0	10798.9	324.1	0.0319	11.4	33.2	–0.2	–10.9	
dg-1	Kamenka	2	476.1	10908.2	29.4	0.4266	16.1	371.0	1.9	–7.2	
dg-2		4	608.0	14459.7	30.0	0.4310	20.2	481.9	2.4	–5.7	
dg-3		10	346.4	3556.1	84.3	0.4125	4.1	42.2	2.4	–14.6	
dg-4		45	437.5	4966.3	172.0	0.0242	2.5	28.8	1.3	–7.4	
dg-5		48	499.4	4115.2	252.3	0.0098	1.97	16.3	1.4	–8.1	
dg-6		63	88.2	1481.7	122.7	0.0042	0.7	12.0	2.5	–6.3	
dg-7		70	201.4	1242.4	104.8	0.0042	1.9	11.8	2.3	–12.6	
dg-8		78	127.1	751.4	276.7	0.0057	0.46	2.7	3.2	–6.4	
dg-9		108	1153.8	1075.8	81.0	0.0047	14.2	13.2	2.4	–5.7	
dg-10		137	77.6	875.9	153.4	0.0075	0.5	5.7	3.2	–7.8	
dg-11		148	118.8	2340.9	290.3	0.0024	0.29	8.0	3.9	–7.1	
dg-12		218	214.4	5042.2	37.1	0.400	5.8	152.0	4.0	–4.4	
dg-13		228	129.5	2236.1	25.6	0.5543	5.0	87.3	4.1	–7.0	
dg-14		320	152.3	2065.6	430.0	0.0203	0.35	4.8	4.8	–5.0	
dg-15		365				0.4812			2.9	–7.6	
dg-16		380	1521.7	4366.2	178.3	0.1997	8.5	24.4	1.1	–12.1	
pt_2-1	Gorbilok	0.5	390.6	3956.6	232.0	0.0042	1.7	17.0	–0.5	–11.3	
pt_2-2		5	242.5	6762.9	121.6	0.0339	2.0	55.6	0.5	–12.4	
pt_2-3		95	107.0	2512.3	291.4	0.0038	0.36	8.6	2.2	–7.9	
pt_2-4		140	275.8	3215.7	203.6	0.0028	1.35	15.8	1.4	–6.8	
sn-1	Kamenka	260				0.4858			2.5	–4.4	
sn-2		280				0.0082			2.2	–6.8	–31.7
sn-3		300	2062.2	6447.9	1071.2	0.0116	1.92	6.0	1.9	–8.8	–33.4
sn-4		380	914.3	874.1	912.0	0.0045	1.0	0.96	3.2	–6.6	–30.1
sn-5		680	3420.9	46703.3	546.9	0.3676	6.25	85.3	–5.0	–7.1	–30.8
sn-6		780	1518.3	24110.2	816.0	0.2095	1.86	29.5	–4.7	–7.2	–30.4
sn-7		810	1693.3	37087.9	551.6	0.233	3.07	67.2	–4.9	–6.2	–30.0
sn-8		830	1569.4	26223.8	577.8	0.221	2.5	45.4	–4.7	–7.8	–31.2
sn-9		850	828.6	26327.3	597.9	0.0223	1.38	44.0	–4.7	–8.1	
sn-10		860	508.9	8821.2	517.6	0.0825	0.98	17.0	–2.8	–7.5	
sn-11	Gorbilok	150	3183.2	7552.6	352.8	0.0039	9.0	21.4	–1.9	–11.0	
sn-12		570	5352.9	49039.2	804.2	0.0318	6.6	60.9	–1.1	–8.7	
sn-13		585	5716.7	27902.8	839.2	0.0169	6.8	33.2	–0.5	–9.3	

Table. (Contd.)

1	2	3	4	5	6	7	8	9	10	11	12
sk-1	Kamenka	70	360.0	10335.9	185.4	0.1170	1.9	55.7	1.0	−5.7	
sk-2		120	449.4	10441.3	205.6	0.1168	2.2	50.8	−0.3	−5.7	
sk-3		123	137.9	1141.4	223.7	0.0087	0.9	5.1	0.5	−7.9	
sk-4		133	241.0	1821.5	257.4	0.0039	0.9	7.0	0.3	−8.2	
sk-5		153	173.8	2967.1	138.0	0.0526	1.3	21.5	0.4	−7.7	
sk-6		163	309.9	14520.2	156.4	0.1355	1.97	92.8	0.9	−4.7	
sk-7		181	51.3	814.8	137.6	0.1024	0.4	5.9	1.5	−7.1	
sk-8		198	33.2	429.3	118.0	0.0112	0.3	3.6	1.5	−6.4	
sk-9		211	516.9	17695.5	90.8	0.2806	5.7	194.8	1.7	−4.2	
sk-10		212	129.8	1105.6	136.7	0.0473	0.95	8.0	1.7	−5.7	
sk-11		219	43.1	642.2	145.6	0.0016	0.3	4.4	1.6	−6.5	
sk-12		240	73.9	361.3	122.4	0.0032	0.6	2.9	1.4	−5.7	
sk-13		345	60.3	345.4	182.1	0.0113	0.3	1.9	1.3	−6.5	
sk-14		350	112.5	835.7	151.1	0.0472	0.7	5.5	1.4	−8.7	
sk-15		440	225.0	1395.3	147.3	0.0039	1.5	9.4	2.6	−7.7	
sk-16a		140				0.0118			4.4	−2.9	
sk-17a		160	134.8	1008.7	356.6	0.0102	0.4	2.8	3.5	−4.0	
sk-18a		163	260.9	1777.8	137.9	0.0107	1.9	12.8	1.9	−4.7	
sk-19a		175	365.9	4093.6	31.7	0.3065	11.5	129.1	3.2	−0.3	
sk-20a		181	288.0	3100.8	51.9	0.3999	5.5	59.7	3.2	−1.3	
sk-21a		184	282.3	3146.4	36.0	0.4570	7.8	87.3	3.8	−1.2	
sk-22a		192	360.0	3546.5	58.4	0.4571	6.2	60.7	3.4	−1.8	
sk-23a		194	366.8	5358.4	40.8	0.3856	8.9	131.0	3.9	−0.6	
sk-24a		202	436.8	5977.5	35.4	0.4500	12.3	168.8	3.7	−0.6	
sk-25a		205	327.2	4555.6	31.8	0.4626	10.2	86.1	3.0	−3.3	
sk-26a		207	344.1	2738.5	32.5	0.4790	10.5	84.2	3.3	−2.5	
sk-27a		212	380.0	4920.0	29.5	0.4737	12.8	166.7	4.3	−0.6	
sk-28a		217	612.6	4826.1	22.5	0.5022	27.2	214.4	2.6	−4.6	
sk-29a		226	794.5	7454.5	32.0	0.4610	24.8	232.9	3.1	−4.9	
sk-30a		245	977.9	7717.6	43.1	0.4421	22.6	179.0	3.3	−4.7	
sk-31a		251	239.4	5904.0	15.8	0.4819	15.1	373.6	2.6	−5.5	
sk-32a		278	385.2	4555.6	26.1	0.4643	14.7	174.5	2.9	−4.6	
sk-33a		284	130.7	3611.0	33.9	0.4937	3.8	106.5	3.7	0.4	
sk-34a		300	190.0	3045.7	35.2	0.4737	5.4	86.5	3.7	−0.6	
sk-35a		314	186.5	4249.1	40.0	0.4920	4.6	106.2	3.6	0.5	
sk-36a		320	168.3	3514.3	38.6	0.4851	4.4	91.0	3.4	−1.0	
sk-37a		334	285.0	7569.2	47.4	0.4918	6.0	159.6	1.9	0.1	
sk-38a		362	132.4	2236.4	37.4	0.4892	3.5	59.8	3.0	−2.1	
sk-39a		373	151.3	2612.4	37.4	0.4921	4.0	69.8	2.3	−1.0	
sk-40a		389	277.0	1609.3	27.9	0.4716	9.9	57.6	2.2	−4.3	
sk-41a		402	358.8	3188.9	21.2	0.5053	16.9	150.3	1.2	−5.3	
sk-42a		412	306.1	5467.7	17.9	0.4757	17.0	305.4	3.1	−6.0	
sk-43a		439	380.0	4555.6	24.5	0.5053	15.5	185.9	2.4	−6.2	
sk-44a		451	276.7	4299.0	23.9	0.5247	11.5	143.7	2.1	−8.1	
sk-45a		461	860.3	11603.8	36.6	0.4872	23.5	317.0	3.2	−2.5	

Table. (Contd.)

1	2	3	4	5	6	7	8	9	10	11	12
sk-46a		491	219.6	4288.1	251.9	0.0226	0.87	17.0	2.9	–5.0	
sk-47a		509	225.3	4004.7	313.8	0.0080	0.7	12.7	2.4	–5.3	
sk-48a		533	2991.1	13396.0	69.7	0.3842	42.9	192.1	2.3	–5.1	
sk-49a		547	5756.4	4384.2	26.2	0.5083	219.6	167.3	3.3	–5.3	
sk-50a		553	706.4	3644.4	19.5	0.5126	20.8	186.8	4.4	–9.2	
sk-51a		575				0.5072			3.7	–6.1	
sk-52a		595				0.5143			1.9	–6.3	
br-1	Gorbilok	3				0.0123			1.6	–7.2	
br-2		15	304.7	3617.6	130.0	0.0037	2.3	27.8	1.1	–12.5	
dd-1	Kamenka	280	135.8	1448.0	31.4	0.4178	4.3	46.1	3.5	–8.8	
dd-2		340	137.6	3454.2	112.8	0.0086	1.2	30.6	4.0	–13.7	
dd-3		380	232.6	4543.8	51.3	0.0682	4.5	88.5	2.8	–12.6	
dd-4		385	341.0	8975.9	70.2	0.1942	4.8	127.8	3.0	–8.0	
dd-5		395	1125.0	7267.4	83.4	0.0346	13.4	87.1	3.2	–8.9	
dd-6		405	162.6	5176.3	53.7	0.5031	3.0	96.3	3.0	–6.7	
dd-7		445	303.1	2570.4	58.6	0.0548	5.2	43.8	2.6	–12.8	
rb-1	Gorbilok	360	527.1	10040.6	36.1	0.3169	14.6	278.8	3.9	–7.8	
rb-2		440	175.7	3618.9	996.8	0.0057	0.17	3.6	4.2	–10.1	
rb-3		445	177.5	2874.2	1070.0	0.0049	0.16	2.7	4.5	–10.0	
rb-4		465	186.7	2215.7	90.6	0.2820	1.9	24.4	1.9	–11.9	
ds-1	Kamenka	100	298.4	6675.3	1750.0	0.0511	0.17	3.8	0.3	–7.7	
ds-2		115	851.6	15546.4	637.2	0.0813	1.3	24.3	0.3	–0.5	
ds-3		121	732.8	10745.3	610.6	0.0856	1.2	17.5	0.5	–8.2	
ds-4		160				0.0821			5.0	–7.8	–25.6
ds-5		185	1431.6	5540.9	914.4	0.0352	1.6	6.0	4.0	–7.4	
ds-6		225	374.9	5645.9	944.1	0.0368	0.39	5.97	3.7	–7.2	
ds-7		275	132.0	2325.6	300.2	0.1066	0.4	7.7	1.0	–6.7	
ds-8		305	644.7	1423.1	458.0	0.0085	1.4	3.1	3.7	–6.8	
ds-9		350	205.7	3571.4	440.8	0.1080	0.46	8.1	5.3	–3.7	
ds-10		475				0.0078			4.1	–6.3	–25.5
ds-11		515	42.3	820.8	152.0	0.0158	0.3	5.4	5.3	–6.3	
ds-12		542	84.0	1046.5	144.5	0.0159	0.6	7.2	5.2	–5.8	
ds-13		545	78.5	2114.2	97.0	0.1397	0.8	21.7	5.0	–2.3	
ds-14		555	73.1	1112.9	115.0	0.0827	0.6	9.7	4.2	–3.8	–22.9
ds-15		560	72.0	2790.7	88.9	0.0732	0.8	31.3	3.0	–7.2	
ds-16		580	71.3	504.1	140.9	0.0068	0.5	3.6	3.7	–5.8	
ds-17		587	118.1	1722.7	162.5	0.0267	0.7	10.6	2.9	–5.7	
ds-18		625	63.1	530.6	137.7	0.0089	0.45	3.9	3.4	–4.7	
ds-19		700	191.2	4578.5	184.0	0.0906	1.0	24.8	2.9	–4.2	
ds-20		718	139.7	1596.7	157.2	0.0204	0.9	10.1	3.9	–6.1	
ds-21		725	180.0	8236.4	105.2	0.2817	1.7	78.2	4.6	0.0	
ds-22		730	58.6	1050.3	169.7	0.0147	0.3	6.2	2.9	–5.6	
ds-23		793	200.0	10335.9	105.2	0.4345	1.9	101.1	3.7	–4.4	
ds-24		835	42.3	957.6	206.2	0.0318	0.2	4.6	2.4	–2.9	
ds-25		850	48.0	1434.1	189.3	0.0107	0.25	7.6	3.3	–3.1	

Table. (Contd.)

1	2	3	4	5	6	7	8	9	10	11	12
ds-26		865				0.0078			4.9	–3.4	–24.5
ds-27		877	36.6	670.0	142.6	0.0071	0.25	4.7	3.2	–3.6	
ds-28		905	64.6	539.3	172.5	0.0079	0.4	3.1	3.5	–6.2	
ds-29		917	60.0	930.2	119.2	0.0134	0.5	7.8	3.3	–7.0	
ds-30		945	549.7	1024.2	317.1	0.0087	1.7	3.2	2.3	–6.0	
ds-31		961	183.3	1058.2	383.3	0.0075	0.47	2.8	4.1	–6.1	
ds-32		1055	50.8	724.2	1482.6	0.0052	0.03	0.48	4.9	–5.2	–24.0
ds-33		1075	222.8	504.1	339.3	0.0068	0.65	1.5	4.6	–3.0	
ds-34		1105	167.3	2771.4	374.5	0.0523	0.45	7.4	4.9	–2.1	–24.8
ds-35		1110	56.3	877.5	191.1	0.0245	0.3	4.6	5.3	–8.6	
ds-36		1118	49.7	624.5	165.0	0.0255	0.3	3.8	3.8	–4.1	
ds-37		1148	51.4	355.7	94.7	0.0255	0.54	3.8	3.6	–4.3	
ds-38		1160	56.9	1293.0	140.0	0.0336	0.4	9.2	2.6	–5.3	
ds-39		1235	32.3	681.7	708.9	0.0031	0.04	0.96	2.6	–5.4	
ds-40		1275	62.0	1764.2	602.6	0.0048	0.1	2.9	3.5	–5.2	
ds-41		1280	47.2	1258.1	724.3	0.0021	0.06	1.7	4.1	–5.5	
ds-42		1400	175.7	5952.4	1781.6	0.0118	0.09	3.3	5.6	–8.3	
ds-43		1450				0.1012			3.1	–7.8	–25.2

Formations: (kd) Korda, (ss) Sosnovyi, (dg) Dzhur, (pt_2) upper Potoskui Subformation, (sn) Shuntarskaya, (sk) Seryi Klyuch (sk-16a–sk-52a designate samples from the Angara section), (br) Borema, (dd) Dadykta, (rb) Rybnaya, (ds) Dashka.

alteration (Veizer, 1983). Carbonate rocks, which experienced the influence of postsedimentation processes and show insignificant changes in C-isotope ratios, are usually characterized by $\delta^{18}O$ values in excess of –10 to –11‰ (Kaufman *et al.*, 1993; Knoll *et al.*, 1995). In the examined samples, they vary from values close to zero to –14‰, and in dolomites, they are usually higher by 2–3‰ than in limestones, which is a consequence of isotope fractionation during dolomitization and of better preservation of the O-isotope system in dolomites (Pokrovskii and Gertsev, 1993; Knoll *et al.*, 1995; Sochava *et al.*, 1996; and others) (table). In general, lower $\delta^{18}O$ values are characteristic of carbonate rocks of the Gorbilok zone, where they experienced a strong catagenetic alteration.

The influence of postsedimentation processes on the C-isotope ratios is estimated using the correlation diagrams for $\delta^{18}O$ and $\delta^{13}C$ values. No correlation between these parameters (Fig. 3) is established for the examined rooks except for an insignificant positive correlation characteristic of carbonates of the Dzhur and Seryi Klyuch formations. Taking into consideration our own and published data of the geochemistry of C and O isotopes, samples of carbonate rocks are divided into three groups. The first group includes rocks with $\delta^{18}O$ values below –10‰, which probably experienced the most significant postsedimentation alteration. The second group of rocks with $\delta^{18}O$ from –10 to –5‰ is characterized by slight alteration, and the third one ($\delta^{18}O$ exceeds –5‰) is represented by unaltered rock varieties. There is a positive correlation between the rock recrystallization degree and $\delta^{18}O$ values: the least recrystallized rocks show higher $\delta^{18}O$ values. For instance, limestones from the Dashka Formation are characterized by a good preservation of primary textural elements and by $\delta^{18}O$ values exceeding –8‰, i.e., they are referred to unaltered or slightly altered carbonate rocks with $\delta^{13}C$ values close to the primary ones. Nonetheless, the direct correlation between the recrystallization degree and $\delta^{18}O$ values observed for some dolomites. So, some strongly recrystallized dolomites almost devoid of primary textural elements from the Dzhur and Dadykta formations show moderate negative $\delta^{18}O$ ratios.

One of the ways to estimate changes in C-isotope ratios in carbonates is based on the difference $\Delta\delta^{13}C$ between $\delta^{13}C_{carb}$ and $\delta^{13}C_{org}$. If the $\Delta\delta^{13}C$ values are 28–30‰, then $\delta^{13}C$ values in carbonate rocks are considered unchanged (Knoll *et al.*, 1986, 1995; Kaufman *et al.*, 1993). Determinations of $^{13}C_{org}$ in some our samples (Timoshina, 1995) indicate that $\Delta\delta^{13}C$ values vary from 35.3 to 25.1‰. The lowest $\Delta\delta^{13}C$ values are recorded in some samples from the Shuntarskaya Formation with significant negative $\delta^{13}C$ values (to –5‰); they probably reflect postsedimentation changes in C_{carb} isotope composition. However, samples with $\Delta\delta^{13}C$ exceeding 32‰ from the same formation can hardly be referred to the strongly altered rock varieties. It seems

that the upper limit of $\Delta\delta^{13}C$ value equal to 30‰ cannot serve as an indicator of considerable changes in the primary C-isotope composition of carbonates (Pell *et al.*, 1993).

Different geochemical criteria are often used to assess the deviation degree of determined $\delta^{13}C$ values from the primary ones: Mn/Sr and Fe/Sr ratios, the presence or absence of covariation between Mn/Sr and $\delta^{13}C$, as well as between Mn/Sr and $\delta^{18}O$ values. These parameters are indicative because, in the course of postsedimentation transformations, carbonate rocks usually become enriched in Mn, Fe, and Rb and impoverished in Sr (Brand and Veizer, 1980; Veizer, 1983; Gorokhov *et al.*, 1995; Kaufman and Knoll, 1995; Sochava *et al.*, 1996; and others). Diagrams of correlation between Mn/Sr and $\delta^{13}C$ and between Fe/Sr and $\delta^{13}C$ show that the $\delta^{13}C$ values depend on Mn/Sr and Fe/Sr ratios in a different manner (Fig. 4). For instance, in limestones from the Dashka Formation, this dependence is ambiguous, being distinct in limestones of the Dzhur Formation. In the considered diagrams, limestone and dolomite fields are clearly separated because of different Sr contents in these rocks. Higher Sr concentrations in limestones are explained by the fact that Sr and Mg ion radii differ from each other more than those of Sr and Ca, and correspondingly, Sr enters the crystal lattice of calcite easier than that of dolomite (Veizer, 1983). In many cases, the enrichment of dolomites in Mn and Fe is also related to the same factor (Veizer *et al.*, 1992). Consequently, the above-mentioned geochemical criteria can reflect, on the one hand, the real postsedimentation alterations of dolomite and, on the other, its crystallochemical properties; in the last case, the primary C_{carb} isotope composition can remain virtually unchanged.

The aforesaid allows different threshold values of element ratios to be used when estimating the extent of dolomite and limestone alteration. Such a differentiated approach was already used for the assessment of $\delta^{13}C$ changes in limestones and dolomites (Sochava *et al.*, 1996). When discriminating samples with disturbed and undisturbed isotopic systems, Sochava and his colleagues used the following Mn/Sr and Fe/Sr values: less than 5 and less than 20 for limestones and less than 10 and less than 40 for dolomites, respectively. Higher than usual critical Fe/Sr values are inferable from diagrams of Mn/Sr and Fe/Sr correlation (Fig. 4). It is seen, for instance, that, when the threshold Mn/Sr value for dolomites is accepted to be less than 10, the corresponding Fe/Sr value should be at least 50.

Summarizing petrographic, isotopic, and geochemical data on carbonate rocks, it should be noted that, despite recrystallization, most of the examined samples preserved $\delta^{13}C$ values close to the primary ones. Deviations of C-isotope ratios from the initial values are most probable for the majority of samples from the Gorbilok zone. Dolomites, isotopically stabilized during early diagenesis, changed afterward their C-isotope composition to an insignificant extent even under strong recrystallization.

RESULTS AND DISCUSSION

The oldest carbonate rocks of the Riphean section in the Yenisei Ridge were sampled from the base of the Korda Formation in the Gorbilok zone. Dolomites are characterized by high Fe and Mn contents and, according to geochemical properties, are classed with the altered rock varieties (table). Petrographic data and O-isotope ratios indicate that only one sample can be considered as slightly altered. The Mn/Sr ratio (11.3) in this sample only slightly exceeds the threshold value (10). The $\delta^{13}C$ values range from –0.1 to –1.9‰, whereas the least altered sample yields –0.8‰. These data indicate that primary C-isotope ratios were apparently close to zero or moderately negative.

Limestone samples from the Sosnovyi Formation of the Gorbilok zone show a moderate recrystallization and, based on geochemical properties, should be considered as representing unaltered rocks (except two samples; table), however, they yield mainly low $\delta^{18}O$ values (less than –10‰). Therefore C-isotope ratios are considered here as unchanged only in a single sample. The $\delta^{13}C$ values vary between –0.5 and 0.7‰. Despite the modified primary C-isotope composition and only slightly contrast absolute values, there is an indistinct trend of $^{13}C/^{12}C$ value increase upward the section; the values are probably lowering at the top of the formation.

Carbonate rocks from the Dzhur Formation are moderately and, sometimes, strongly recrystallized. According to geochemical properties, limestone samples are represented by both altered and unaltered varieties; dolomites are unaltered (Table). The $\delta^{18}O$ values usually exceed –8‰, and only three samples show values below –10‰. There is a slight positive correlation between $\delta^{18}O$ and $\delta^{13}C$ values. The $\delta^{13}C$ values vary from 1.1 to 4.8‰. Higher in the section, C-isotope composition becomes heavier, from 1.3 to 4.0–4.8‰ and shows insignificant variations; the transition to the Shuntarskaya Formation corresponds to $\delta^{13}C$ decrease. It is important that limestone and dolomite samples corresponding in $\delta^{18}O$, Fe/Sr, and Mn/Sr values, as well as in the degree of recrystallization, to altered rock varieties fall within a general trend recording the relative stability of the C-isotope system under postsedimentary alteration.

Limestones of the upper Potoskui Subformation (Dzhur Formation) of the Gorbilok zone are recrystallized and characterized by low (less than –10‰) and moderate (from –6.8 to –7.9‰) $\delta^{18}O$ values. The Fe and Mn contents in the carbonate phase are relatively low; therefore, only one sample should be considered, by geochemical criteria, as altered, but, according to combined geochemical and isotopic properties, two samples from the lower carbonate member with $\delta^{13}C$

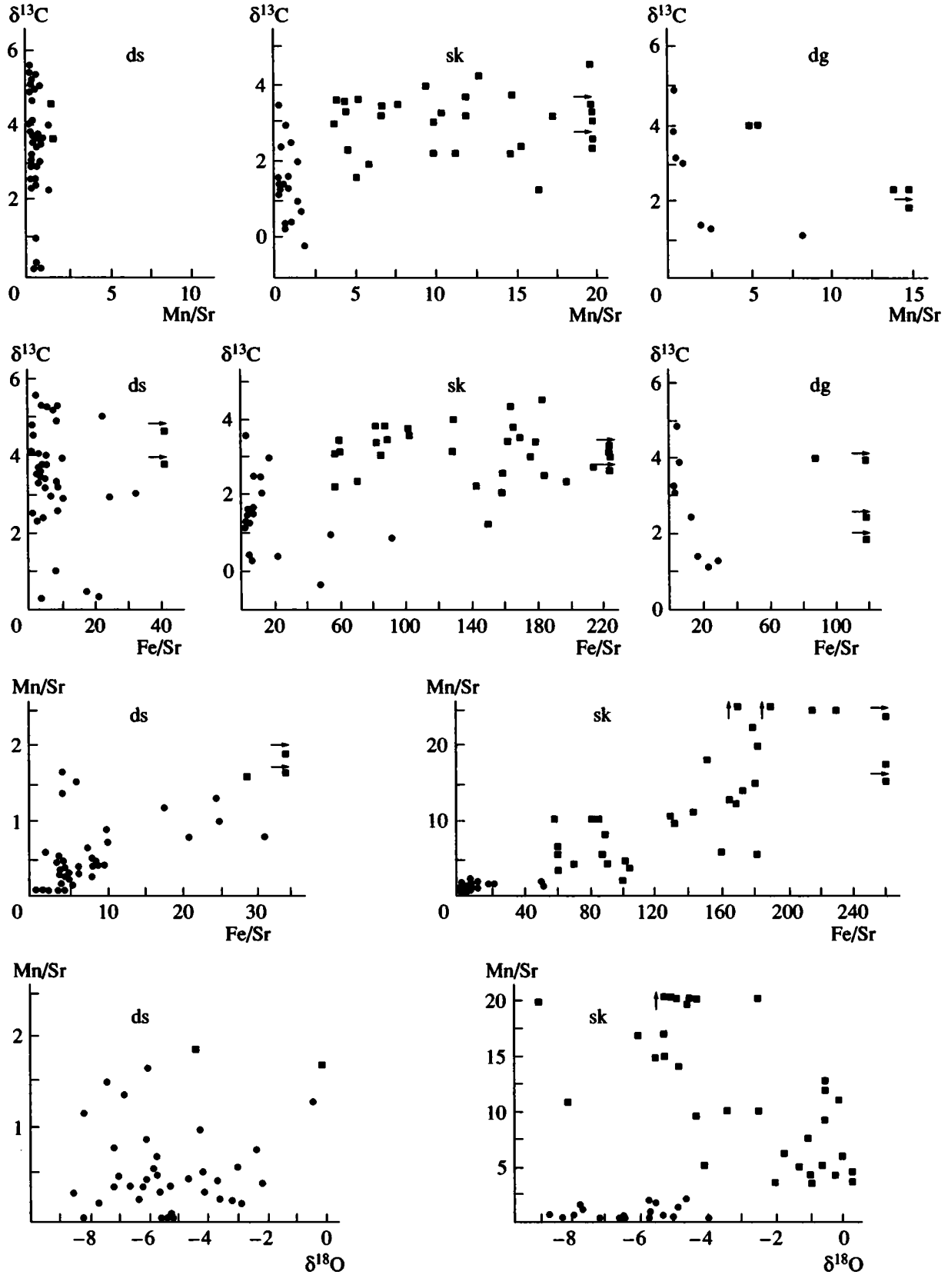


Fig. 4. Isotopic-geochemical cross-plot diagrams for carbonate rocks from the Kamenka zone of the Yenisei Ridge. For legend see Figs. 1, 3.

values of –0.5 and 0.5‰ are referred to altered rocks. Samples from the upper carbonate member with $\delta^{13}C$ values of 1.4 to 2.2‰ are considered as representing unaltered rocks. These data confirm the proposed correlation between deposits of the Dzhur Formation of the Kamenka zone and the upper Potoskui Subformation of the Gorbilok zone, according to which the Potoskui Subformation in the latter zone includes the lateral analogues of only the lower carbonate strata of the Dzhur Formation (Fig. 2).

Carbonate rocks of the Shuntarskaya Formation are moderately recrystallized and yield $\delta^{18}O$ values exceeding –9‰. The $\delta^{18}O$ and $\delta^{13}C$ values show no correlation. In these parameters, C-isotope composition is probably similar to the primary one. Two intervals with different $\delta^{13}C$ values are distinguished in the section: in the lower interval, $\delta^{13}C$ values range from 1.9 to 3.2‰, whereas in the upper one, from –2.8 to –5.0‰. In both cases, the rocks are characterized by an elevated (1.5–3.95‰) organic matter content with low $\delta^{13}C$ values ranging from –30.0 to –33.4‰ (Timoshina, 1995, Kontorovich *et al.*, 1996). The $\Delta\delta^{13}C$ value for samples with positive $\delta^{13}C$ values is 33–35‰, whereas that for samples with light C_{carb} isotope composition ranges from –25.1 to –26.5‰. As noted above, the Neoproterozoic unaltered carbonates are characterized by $\Delta\delta^{13}C$ values of 28–30‰ (Kaufman *et al.*, 1993; Knoll *et al.*, 1995), although other researchers believe that the range of 26–29‰ is more reliable and assume variations from 24 to 34‰ (Schidlowski and Aharon, 1992). Nonetheless, if the $\delta^{13}C$ value of 27–28‰ is accepted as representing the threshold one, then the postsedimentary decrease of heavy carbon in C-isotope composition of carbonate rocks of the Shuntarskaya Formation should be assumed.

Dolomitic silty limestones and limy silty dolomites from the interval with the low $\delta^{13}C$ values are characterized by elevated concentrations of Fe in the carbonate phase and, according to Fe/Sr values, they are mainly referred to altered rocks (table). Nevertheless, in one limestone–dolomite sample (sn-6), the Fe/Sr value only slightly exceeds the threshold one and, thus, the sample can represent the slightly altered rock, whereas another limestone sample (sn-10) with $\delta^{13}C$ value of –2.8‰ is considered, based on isotopic and geochemical properties, as representing the unaltered rock.

The results of lithological analysis suggest that Shuntarskaya carbonate turbidites and hemipelagic rocks were formed in the moderately deep stratified back-arc basin (Khabarov, 1994, 1995; Kontorovich *et al.*, 1996), whose bottom and interstitial waters were enriched in carbonic acid with the light C-isotope composition that resulted from bacterial decomposition of primary organic matter. Therefore, the early cementation of sediments was accompanied by the formation of carbonate minerals sometimes occurring in intergranular spaces and these minerals are characterized by a light C-isotope composition lower than the C-isotope ratios in rocks as a whole. This phenomenon is exemplified in many works (Aharon and Liew, 1992; Derry *et al.*, 1992; Ripperdan, 1994; Pokrovskii, 1996; Tobin *et al.*, 1996; and others). C_{carb} isotope composition probably became lighter also at the later diagenetic stages, as is evident from the elevated Fe/Sr and lowered $\Delta\delta^{13}C$ values in some samples. However, the phenomenon in question cannot explain the occurrence of carbonate rocks in two stratigraphic intervals with similar sedimentation and diagenetic histories but differing from each other in the $\delta^{13}C$ value by 7–8‰. These differences most likely reflect secular variations of carbon isotope composition in world oceans. Taking into account all the possible asumptions, it is believed that the primary negative $\delta^{13}C$ values for carbonates were close to –2 or –3‰.

Three samples of carbonaceous (1.5–3.5% of C_{org}) limestones from the Shuntarskaya Formation of the Gorbilok zone were analyzed. They are characterized by elevated levels of Fe and Mn and, in accordance with geochemical properties, represented the altered rocks. The limestone sample (sn-11) yields the lesser $\delta^{18}O$ and $\delta^{13}C$ values (–11.0 and –1.9‰, respectively). Two other samples with $\delta^{13}C$ values of –0.5 and –1.1‰ were collected from deposits that represent a lateral analog of the lower part of the Seryi Klyuch Formation, but show lower $\delta^{13}C$ values. This fact is easily explained both by the above-mentioned early diagenetic changes in C-isotope composition and by the isotopic exchange between carbonate and organic carbon in the course of late catagenesis (early stage of greenschist metamorphism).

Two exposures of the Seryi Klyuch Formation were examined: the Angara (dolomites) and Kamenka (limestones) sections. Early diagenetic dolomites are characterized by either weak, or, rarely, strong recrystallization. The $\delta^{18}O$ values vary from –9.0 to 0.4‰. There is a correlation between $\delta^{18}O$ and $\delta^{13}C$ values (Fig. 3). Dolomites show the relatively high Fe, Mn, and low Sr contents (Table). Therefore, most of the samples yield high Mn/Sr and, particularly, Fe/Sr values, but no correlation between these element ratios and $\delta^{13}C$ values is observed (Fig. 4). In this connection, only Mn/Sr values were accepted as a criterion for discriminating between altered and unaltered dolomites. Nonetheless, even in this case, $\delta^{13}C$ values in unaltered samples often appear to be lower than those in altered varieties and range from 1.2 to 4.4‰ (Fig. 5b). Limestones sampled from the lower and upper portions of the Angara section are referred to unaltered rocks with $\delta^{13}C$ values of 1.9–4.4‰.

Limestones of the Kamenka section are moderately recrystallized and variably dolomitized; their $\delta^{18}O$ values vary from –4.0 to –8.7‰. There is a slight positive correlation between $\delta^{18}O$ and $\delta^{13}C$ values. The Mn/Sr and Fe/Sr values are relatively low, but the Fe content sharply increases in strongly dolomitized varieties, and,

accordingly, the Fe/Sr value also grows. The relevant samples are referred to as altered rocks, although in the $\delta^{13}C$ value they do not differ from slightly altered varieties (Table, Fig. 5b). The C-isotope ratios vary from –0.3 to 2.6‰. Higher in the section, $\delta^{13}C$ values increase; slightly lower values are registered in the lower and middle portions of the section. The similar trend of C-isotope ratio variations is also recorded in the dolomite section.

Comparison of $\delta^{13}C$ behavior in these sections indicates that this parameter differs by 1.0–1.8‰. This fact can be explained by the different age of deposits, by a strong impact of postsedimentary processes, and by the dependence of C-isotope ratios on sedimentation environments.

The stratigraphic position of the Kamenka section, where the lower and upper boundaries of the considered formation are distinct, is well defined. This is less true for the Angara stratotype section of the Seryi Klyuch Formation, where the formation boundaries are poorly exposed and disturbed by faults. Nonetheless, there are no grounds at present to call into question a similar stratigraphic position of the Angara and Kamenka sections.

A substantially lighter C-isotope composition in response to postsedimentary processes can be assumed for rocks of the limestone section, but this is inconsistent with petrographic, isotopic, and geochemical data. The data evidence that carbonate rocks from the dolomite section are altered even more than those from the limestone section. Consequently, an assumption that more a significant postsedimentary transformation of rocks from the Kamenka section are accompanied by the shift of the C-isotope parameters by 1.0–1.8‰ is not confirmed by the factual data. Simultaneously, the discrepancy between $\delta^{13}C$ values in both sections can be related to the specific distribution of the $\delta^{13}C$ value in CO_2 of seawater in different sedimentation environments.

It was noted previously that deposits of the dolomite section were formed within the restricted and moderately restricted shelf, whereas those of the limestone section accumulated mainly in the basin and slope settings, when the shelf margin was occupied by stromatolitic reefs (Khabarov, 1994; Khabarov *et al.*, 1998). The available data on C-isotope composition of the dissolved CO_2 in deep and surface waters of modern sea basins (Broecker, 1982; Aharon and Liew, 1992; Schidlowski and Aharon, 1992; Derry *et al.*, 1996) indicate that the respective values differ by 1.2–2.6‰. It can be assumed that the differentiation of C-isotope composition in carbonic acid of deep and surface waters was also characteristic of the Late Precambrian basins, particularly of those with a stratified water column. Stratigraphic–lithological studies (Khabarov, 1994; Khabarov *et al.*, 1998) indicate that during the accumulation period of the lower part of the Seryi Klyuch Formation, the stratified basin inherited from the Shuntarskaya time continued to develop in the Gorbilok and, partly, in the Kamenka zones (Figs. 2, 5). Accordingly, the participation of the isotopically relatively light bottom and interstitial waters in the early cementation of sediments, accumulated in deep-water settings and derived from the shelf, might result in the relative increase of $\delta^{13}C$ values in carbonates. Later, during the destruction of water stratification in the basin and as a result of upwelling of deep, isotopically light waters and their interaction with slope carbonate sediments, the shift of $\delta^{13}C$ values in carbonates toward the lower values could also be possible. On the other hand, evaporitization could result in the heavier C-isotope composition in carbonates of the restricted shelf (Chafetz and Rush, 1995; Gill *et al.*, 1995; Akhmedov *et al.*, 1996). Intervals with the relatively high (up to positive) $\delta^{18}O$ values confirm this assumption. Thus, the difference in $\delta^{13}C$ values in the limestone and dolomite sections of the Seryi Klyuch Formation is probably related to differentiation of isotopic composition of the bottom and interstitial waters in the plain/slope and restricted shelf settings of the basin.

Two limestone samples were examined from the base of the Borema Formation (the lateral analog of the upper part of the Seryi Klyuch Formation) in the Gorbilok zone. The lower sample with $\delta^{13}C$ and $\delta^{18}O$ values of 1.6 and –7.2‰, respectively, is considered as representing the unaltered rock, whereas the upper one with $\delta^{13}C$ and $\delta^{18}O$ values of 1.1 and –12.5‰, respectively, is referred to the altered variety. The latter is also characterized by the elevated (27.8) Fe/Sr value.

Carbonate rocks of the Dadykta Formation are partly or, sometimes (dolomites), strongly recrystallized. Their $\delta^{18}O$ values vary from –6.7 to –13.7‰. The Fe/Sr ratios in all samples and Mn/Sr ratios in two samples exceed the threshold value (table). The elevated geochemical ratios correlate with the low Sr contents (31.4–112.8, usually less than 80 ppm) in dolomitized limestones and in dolomites. By all criteria, these rocks are estimated as representing the altered varieties, however, no correlation between $\delta^{18}O$ and $\delta^{13}C$ values is observed (Fig. 3), and $\delta^{13}C$ values ranging from 2.6 to 4.0‰ may be close to the primary ones. Rocks from the middle part of the formation show a insignificantly lighter C-isotope composition.

The C- and O-isotope ratios for the upper part of the Rybnaya Formation in the Gorbilok zone were determined in four samples. Two limestone samples represent, as is evident from their geochemical properties, the unaltered rocks and are characterized by $\delta^{13}C$ values of 4.2 and 4.5‰. The close $\delta^{13}C$ value (3.9‰) is recorded in the dolomite sample from the lower part of the section, although, judging from geochemical parameters, this sample should be considered as being altered. The dolomite sample from the top part of the section ($\delta^{13}C$ 1.9‰) seems also to be altered.

Carbonate rocks of the Dashka Formation are slightly recrystallized. The $\delta^{18}O$ values exceed –8‰,

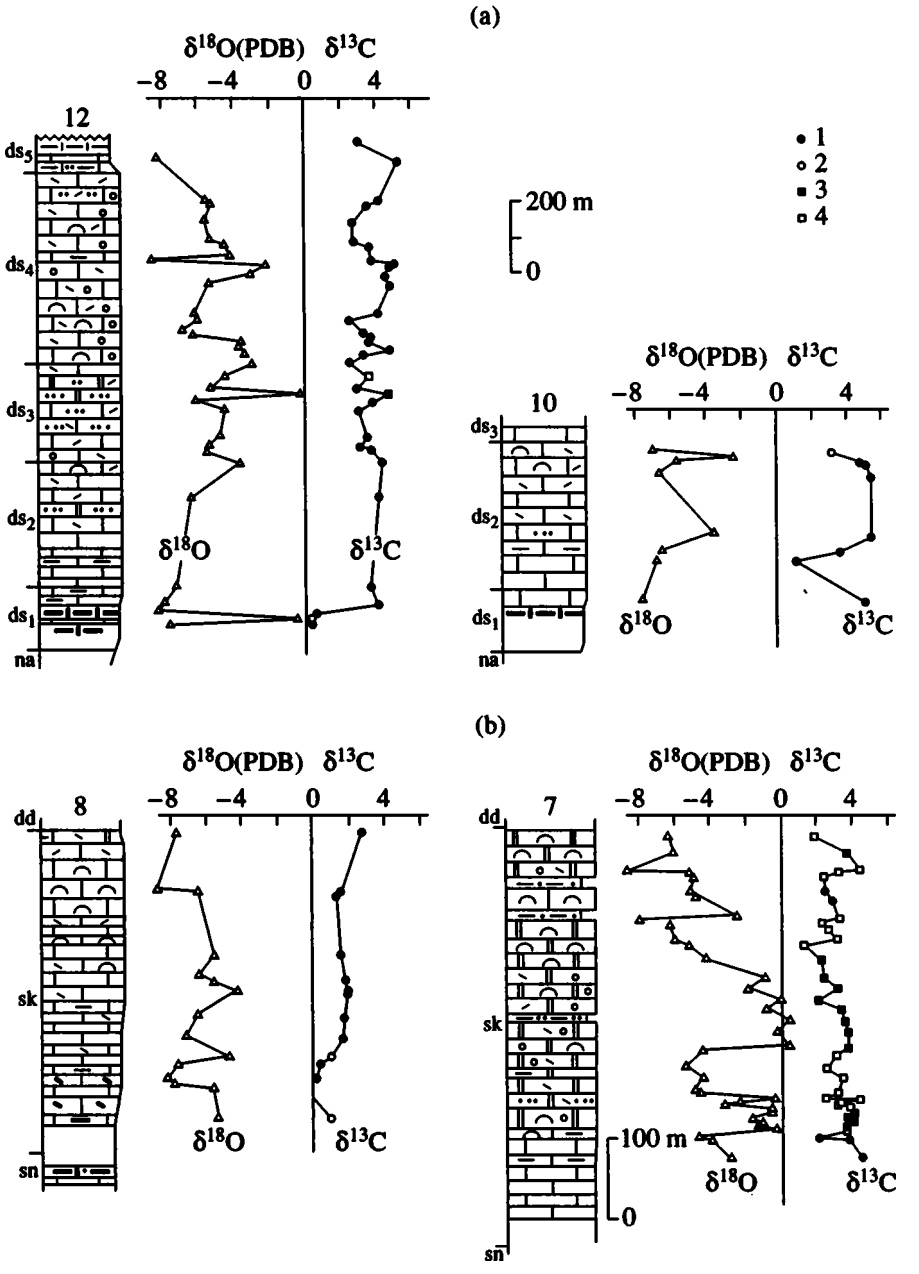


Fig. 5. Variations of $\delta^{18}O$ and $\delta^{13}C$ values in carbonate rocks from the Dashka (a) and Seryi Klyuch (b) formations. (1–2) limestones: (1) unaltered, (2) altered; (3–4) dolomites: (3) unaltered, (4) altered. Other symbols as in Figs. 1, 2.

and no distinct correlation exists between $\delta^{13}C$, $\delta^{18}O$, Mn/Sr, and Fe/Sr values (Figs. 3, 4). The $\Delta\delta^{13}C$ values range from 27.1 to 30.6‰ and indicates that the C-isotope parameters are close to the primary ones. However, in some samples from the basal and middle parts of the section, Fe/Sr ratios significantly exceed the threshold value characteristic of limestones and dolomites (Table). The $\delta^{13}C$ values vary from 0.3 to 5.6‰ averaging about 3.7‰. The lowest $\delta^{13}C$ values are observed at the base of the section; the main part of the section is characterized by an alternation of intervals with moderately high (4–5‰) and relatively low (2.5–3.0‰) values (Fig. 5a). The lighter C-isotope composition in samples from the basal part of the formation is a result of secular variations of $\delta^{13}C$ values in seawater, but can be partly related to the slightly elevated (to 0.25%) content of organic matter and to the influence of the isotopically light CO_2 produced by bacteria.

Summarizing all the data obtained for the Riphean section of the Yenisei Ridge, the following main changes in $\delta^{13}C$ values should be stressed. Carbonate rocks from the lower part of the Korda Formation, whose age is estimated to be 1450–1350 Ma, are char-

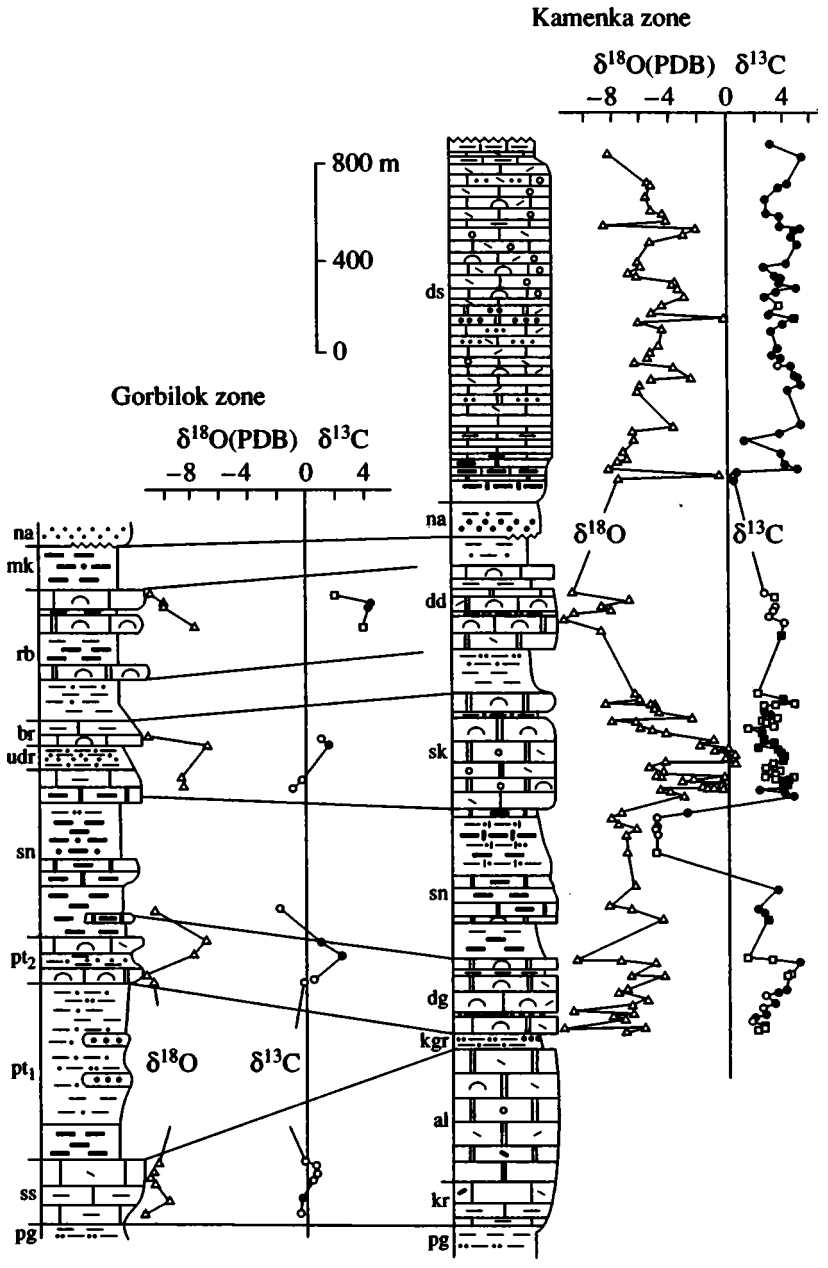


Fig. 6. Variations $\delta^{18}O$ and $\delta^{13}C$ values in carbonate rocks from eastern zones of the Yenisei Ridge. For legend see Figs. 1, 2, 5.

acterized by $\delta^{13}C$ values close to zero. Similar $\delta^{13}C$ values are also registered in the carbonate turbidites of the Sosnovyi Formation 1100–1000 Ma old. Despite a slight difference in absolute $\delta^{13}C$ values, there is an indistinct ascending trend of changes in the $^{13}C/^{12}C$ ratio upward the section and the probable lowering of this parameter at the top of the formation (Fig. 6). A similar trend is established for the Alad'ino Formation (the section on the right bank of the Angara River, opposite the Manzya Settlement). At the last site, the maximum $\delta^{13}C$ value amounts to 2.7‰, and it decreases to 1.9‰ at the top of the formation and to 0.3‰ in the upper part of the Kartochki Formation.

Deposits dated as 1000 to 850–820 Ma old are characterized by alternating durable intervals with moderately high (2.5–4.5‰, maximum to 5.6‰) positive $\delta^{13}C$ values and relatively short intervals with lower values. The Dzhur deposits show the distinct ascending trend of changes in $\delta^{13}C$ values from 1.6–1.9 to 4.8‰. The trend is complicated by insignificant fluctuations throughout the section and also by a minimum at the top of the formation. Based on C-isotope data, the Shuntarskaya Formation can be subdivided into two parts. The lower part is characterized by the ascending trend of $\delta^{13}C$ values from 1.9 to 3.2‰, whereas the upper one is marked by a sharp negative peak (–4.7 to

–5.0‰) followed by the parameter increase to –2.8‰ at the top of the formation. The late Shuntarskaya negative peak is undoubtedly related to secular variations of the C-isotope composition in seawater. Taking into consideration the possible early and late diagenetic impact on the C-isotope systems, the primary $\delta^{13}C$ values probably did not exceed –2.0 to –3.0‰ during these events. The peak interval is characterized by an elevated influx of siliciclastic material to the deep-water basin and by a subsequent progradation of shelf carbonates. The event probably reflects a eustatic sea-level fall.

The ascending trend of $\delta^{13}C$ value variations in the uppermost part of the Shuntarskaya Formation also continues to the lower part of the Seryi Klyuch Formation. In the Angara section of the latter, $\delta^{13}C$ values rapidly increase to 4.4‰ and then rapidly fall to 1.9‰. Higher in the section, $\delta^{13}C$ values mostly vary from 3.0 to 4.4‰, sometimes decreasing to 1.2–2.6‰. Carbonate subformations of the Dadykta (Kamenka zone) and Rybnaya (Gorbilok zone) formations are characterized by relatively uniform $\delta^{13}C$ values of 3.0–4.5‰; these parameters display slight fluctuations and probably fall at the top of the formation (Fig. 6). The lower part of the Dashka Formation is marked by the ascending trend of $\delta^{13}C$ variations from 0.3 to 5.0‰ followed by their fall to 1.0‰ and subsequent increase to 5.3‰. The largest portion of the Dashka Formation represents an alternation of durable intervals with moderately high $\delta^{13}C$ values (3.5–5.6‰) and short intervals with its lower values (2.3–3.0‰).

CONCLUSION

The paper presents the results of the study on the carbon and oxygen isotope composition in Riphean rocks from the eastern zones of the Yenisei Ridge, which provide additional information for specifying the model curve of $\delta^{13}C$ variations in the Riphean. The reliability of the obtained $\delta^{13}C$ values is substantiated by petrographic, geochemical, and isotopic data, which indicate that the best examined samples from the Kamenka zone show no significant postsedimentary distortions in the C-isotope systems and can be used for chemostratigraphic purposes. This is also true for samples classed with the altered rocks, but characterized by the $\delta^{13}C$ values following the general trend. Simultaneously, samples from the Gorbilok zone affected by late catagenesis and early metamorphism mainly show C-isotope parameters different from the primary ones. The exception is a series of samples collected from the thick turbidite members with minimal amounts of siliciclastic material and organic matter having the light C-isotope composition. It should be noted that coeval carbonate rocks devoid of the postsedimentary imprint, but formed in sedimentation settings, where C-isotope composition in bottom and interstitial waters was differentiated, can also differ in absolute $\delta^{13}C$ values. This conclusion emphasizes an importance of sedimentological reconstructions for the inter- and intrabasin correlation based on C-isotope data.

Comparison of obtained absolute values and trends in $\delta^{13}C$ variations with similar parameters recorded in the Turukhansk section of the western margin of the Siberian platform (Knoll *et al.*, 1995), which, in opinion of many researchers (Semikhatov and Serebryakov, 1983; Shenfil', 1991; and others), correlates with the carbonate section of the Yenisei Ridge, indicates that the Sosnovyi Formation, on the one hand, and Linok and Sukhaya Tunguska formations, on the other, are characterized by close $\delta^{13}C$ values. The overlying Derevnya and Dzhur formations show a similar trend (an increase of $\delta^{13}C$ values upsection), but absolute values of C-isotope ratios in their carbonates are substantially different. In the Dzhur Formation, these values amount to 4.8‰, whereas in the Derevnya Formation, they do not exceed 2.6‰. The detailed correlation of overlying carbonate deposits on the basis of C-isotope data, presents certain difficulties, despite some common trends established here (prevalence of intervals with relatively high $\delta^{13}C$ values). The early Shorikha negative peak of C-isotope parameters probably corresponds to the similar late Shuntarskaya negative peak, whereas that of the late Shorikha time correlates with the early Dashka one, although their amplitudes sharply differ from each other.

The analysis of curves demonstrating $\delta^{13}C$ variations in rocks of the Yenisei Ridge and neighboring Baikit arch (Khabarov *et al.*, 1996) reveals that they are substantially different in the general evolutionary trend and absolute $\delta^{13}C$ values. In the Yenisei Ridge, C-isotope ratios usually are as high as 3.5–4.5‰ (maximum 5.6‰), whereas in the Baikit arch, they rarely exceed 1.0‰ amounting to 2.7‰ at the top of the section. These data are inconsistent with the available correlation scheme, according to which the substantially carbonate part of the section of the Yenisei Ridge (from the Kartochki Formation upward) is coeval to the carbonate section of the Baikit arch (Surkov *et al.*, 1996). However, the data confirm the older, mainly Early–Middle Riphean (1500–1450 to 950 Ma) age of carbonate deposits in the Baikit arch that is suggested in some works (Khabarov *et al.*, 1996; Kontorovich *et al.*, 1996). Based on the C-isotope parameters, the correlation may be as follows. The lower part of the Korda Formation with the moderate negative $\delta^{13}C$ values corresponds to the lower part of the carbonate section in the Baikit arch, where carbonates yield negative and low positive values. The Sosnovyi Formation, characterized by $\delta^{13}C$ values of –0.5 to +0.7‰, correlates with deposits from the upper part of the Baikit section, which are 1080–1000 Ma old and show $\delta^{13}C$ values of –0.7 to 1.7‰. The positive trend (from –0.5 to 2.7‰) registered in the uppermost part of the Baikit section corresponds to a similar trend in the lower part of the Dzhur Formation.

Comparison of our C-isotope data with those obtained for other regions reveals that the age of the examined deposits (except the Korda Formation) is not older than 1200 Ma, because the older Riphean carbonate rocks are mostly characterized by $\delta^{13}C$ values close to zero (Knoll *et al.*, 1995), although probably not universally (Ghazban *et al.*, 1992; Strauss *et al.*, 1992; Khabarov *et al.*, 1996). In general, $\delta^{13}C$ values from the upper Sukhoi Pit, Tungusik, and Oslyanka groups suggest that they can be correlated with deposits 1100–850 Ma old, being in agreement with available isotopic–geochronological dates. Simultaneously, the Dadykta and Dashka formations with the elevated $\delta^{13}C$ values (4–5‰) may also correspond to younger (less than 850 Ma old) deposits displaying isotopic parameters (Knoll *et al.*, 1986, 1995), which are characteristic of the early phase of the Late Proterozoic interval of the positive $\delta^{13}C$ values.

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The Late Quaternary Hydrological History of the Sea of Okhotsk as Evidenced by Foraminifers

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Abstract—Benthic foraminifers from sediments, which have been obtained at the slope of the Kuril island arc in the southeastern part of the Sea of Okhotsk (Core V34-98, depth of 1175 m) and dated by using the AMS radiocarbon technique, were studied. Taxonomic composition, species diversity and abundance, as well as changes of these characteristics were used to reconstruct main hydrological events in the basin during the late Pleistocene and Holocene. It is shown that during the last 17000 years, the bottom of Sea of Okhotsk was successively populated by three benthic foraminiferal assemblages different from each other by dominant species, such as the *Eponides weddellensis, Brizalian spissa*, and *Bolivina seminuda* assemblages. The assemblages characterize particular hydrologic environments and correspond to three main periods of the late Pleistocene–Holocene history of the basin evolution, or to three climatostratigraphic intervals: the terminal stage of the last glaciation (17–12.5 ka), deglaciation stage (12.5–8.5 ka), and last interglacial, or postdeglaciation period (last 8 ka). Changes in the structure of benthic foraminiferal assemblages are consistent with data on the oxygen isotope composition of their carbonate shells.

Key words: *benthic foraminifers, oxygen isotope composition, radiocarbon dating, late Pleistocene, Holocene, glaciation, deglaciation, postdeglaciation, Sea of Okhotsk*

INTRODUCTION

Studies on the peculiarities of the hydrological regime in seas and oceans, as well as the regime transformation through the Quaternary glacial and interglacial epochs have recently been making progress in connection with the need to elaborate the models of climate evolution during the Quaternary and, based on the latter, to predict climatic changes in the future. The creation of reliable oceanographic and climatic models depends, first and foremost on the resolution degree of the performed studies. The most suitable objects for this purpose are marginal seas, where the high sedimentation rates allow the short-term events to be registered. The Sea of Okhotsk, where sedimentation rates were extremely high during the Quaternary (Bezrukov and Lisitsyn, 1957), represents precisely such an object. In addition, owing to the high-latitude position of this marginal basin, oceanic and climatic events that are recorded in its sedimentary cover, are manifested there most distinctly. The importance of studies in the Sea of Okhotsk is also stressed by the fact that the Quaternary climatic and oceanic evolution of the sea was closely related to hydrological processes in the northwestern Pacific. Thus, materials on the Sea of Okhotsk can also shed light on the history of water circulation in the Pacific.

With respect to paleoceanographic history, the Sea of Okhotsk remains poorly studied. By now, there is a limited number of publications dedicated to the Late Quaternary evolution of the basin. To reconstruct the Late Quaternary hydrological evolution of the Sea of Okhotsk, many researchers previously used either lithological properties of sediments (Bezrukov and Lisitsyn, 1957) or enclosed remnants of calcareous and siliceous planktonic microorganisms, in particular planktonic foraminifers and radiolarians (Kurihara, 1982; Keigwin, 1989; Morley *et al.*, 1991; Gorbarenko, 1991; Gorbarenko *et al.*, 1998; Chekhovskaya and Basov, 1999). At the same time, benthic foraminifers, whose taxonomic composition and changes in assemblages reliably reflect properties of water masses; thus, circulation patterns in a sea basin during different time periods were virtually never used for paleoceanographic interpretations in the basin under consideration. Few available publications on benthic foraminifers from the Sea of Okhotsk deal primarily with their distribution in the recent bottom sediments (Saidova, 1961, 1994, 1997; Basov and Khusid, 1983). Only recently, there appeared the first publication that considers, along with planktonic foraminifers, changes in the composition of benthic foraminiferal assemblages in Pleistocene–Holocene sediments of the central part of the Sea of Okhotsk and, based on the latter, discusses paleohydrological and paleoclimatic events during the last 60 ka (Belyaeva and Burmistrova, 1997). These researchers noted, in particular, several warming epi-

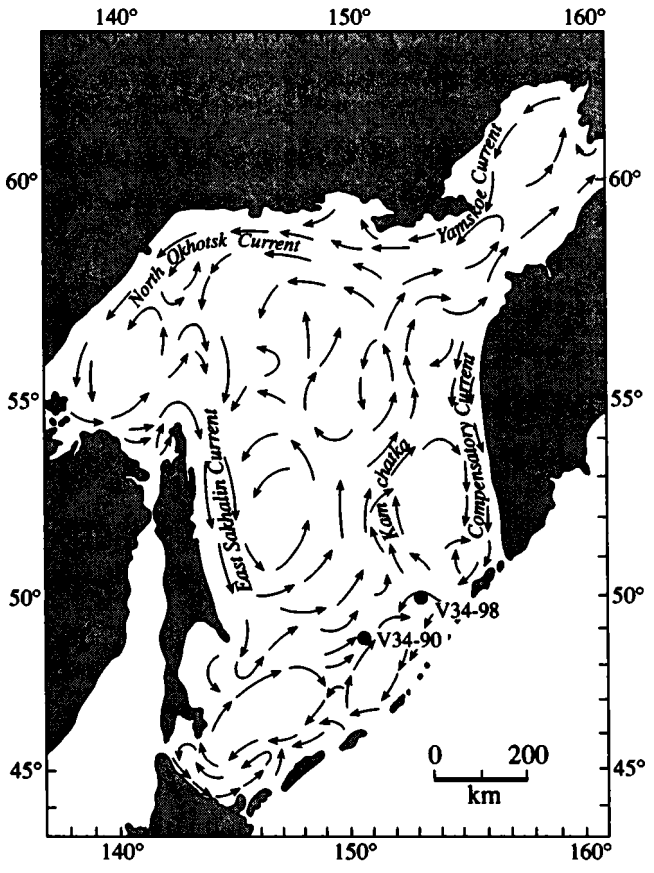

Fig. 1. Generalized scheme of surface circulation in the Sea of Okhotsk and location of Cores V34-98 and V34-90.

sodes of surface and deep waters during this time period, which were most prominent about 11–13 and 10 ka ago.

The paper attempts to reconstruct the hydrologic evolution of the considered marginal sea basin during the last glaciation, deglaciation, and postdeglaciation periods and to date main paleoceanographic and paleoclimatic events based on the analysis of benthic foraminiferal assemblages and their structure in the upper Pleistocene–Holocene sediments dated by the radiocarbon method.

MATERIALS AND METHODS

Sediment samples for the foraminiferal analysis were collected from Core V34-98 (coordinates: 50°006.6N, 153°012.0E; depth of 1175 m) taken from the Kuril island arc slope in the southeastern part of the Sea of Okhotsk during Leg 34 of R/V *Vulkanolog* in 1982 (Fig. 1). The section is 330 cm long. It consists of three members composed of gray silt (interval of 250–330 cm), greenish gray pelitic silt (interval of 135–250 cm), and green diatom ooze (upper 135 cm). Sediments of the lower member are characterized by minimal $CaCO_3$ and C_{org} contents (less than 2 and 0.7–1.2%, respectively). Upsection, the $CaCO_3$ content ranges from 3–4 to 5–6‰, amounting in some layers to 7–8% or more. The C_{org} content varies from 0.5–1.0% to 1.5–2.0% gradually increasing upward the section (Fig. 2).

Samples were collected from a 1-cm-long interval with the step of 5 cm. After the standard laboratory treatment, the fraction >0.5 mm of the dried residue was studied under binoculars to determine the taxonomic composition of benthic foraminifers and calculate their total abundance per 1 g of dry sediments and the relative content (%) of every species in a particular sample. In total, 67 samples were studied.

Radiocarbon age determinations were performed in the Lawrence Livermore National Laboratory (California, USA) using the AMS technique; the sample was 2–5 mg. Radiocarbon dates were obtained for the levels of 294, 219, 204, 99, and 45 cm (15070, 11540, 9340, 4850, and 3470 years, respectively). The radiocarbon age is reduced for the reservoir-corrected factor of the Sea of Okhotsk by subtracting 1000 years (Southon *et al.*, 1993). The sediment age between radiocarbon dates was determined by interpolation. Data on the oxygen isotope composition in shells of benthic (*Uvigerina* spp.) and planktonic (*Neogloboquadrina pachyderma* sin.) foraminifers from the studied core were donated by S.A. Gorbarenko (Pacific Oceanographic Institute, Far East Division, Russian Academy of Sciences).

CHARACTERISTICS OF THE STUDIED AREA

The Sea of Okhotsk represents a semiclosed marginal sea basin connected via the Tatar and Soya straits with the Sea of Japan and via Kuril straits with the Pacific. Deepest of the latter are the Kruzenshtern and Bussol' straits (1870 and 2300 m deep, respectively) located near the studied area in the northern part of the Kuril island chain. They serve as pathways for Pacific intermediate waters, which enter the Sea of Okhotsk via the former and return back to the Pacific via the latter.

The greater part of the Sea of Okhotsk is covered by ice for many months; an exception is its southeastern part, free of ice all year round. This determines a peculiar hydrological structure of the basin, which is characterized by a distinct temperature gradient between surface and subsurface waters during the summer season when a depth interval of 30–40 to 150–200 m is occupied by the so-called Okhotsk-Sea water mass. This water mass, having temperatures close to zero, is formed during the winter season and distributed throughout the entire basin (Moroshkin, 1966). In summer, the temperature of surface waters increases to 10–12°C. The Okhotsk-Sea water mass is underlain by the warm Pacific waters, which enter the basin via the Kruzenshtern Strait and spread further northward, filling the deep part of the basin. At the depth interval of 750–1500 m, these waters are characterized by the maximum temperature (2.0–2.5°C), minimal content of

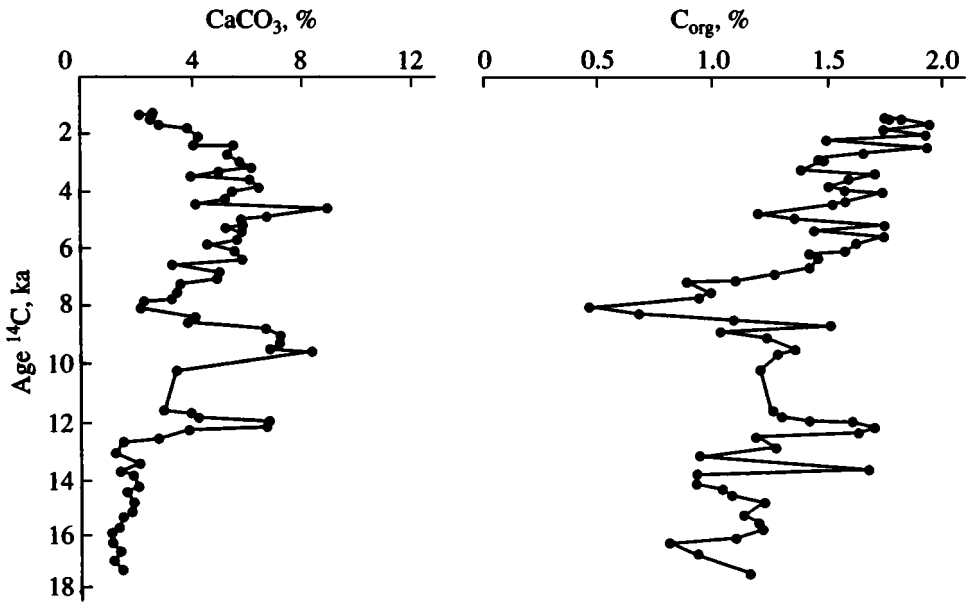

Fig. 2. Variations of the $CaCO_3$ and C_{org} contents (%) in sediments of the core V34-98.

dissolved O_2 (less than 0.5 ml/l), and salinity range between 34.4 and 34.7‰. In the deep Kuril Basin, the water temperature decreases to 1.8°C.

The flow of intermediate Pacific waters is recorded at the bottom by sediments with elevated (to 6–8%) $CaCO_3$ contents (Bezrukov, 1960) and a high abundance of benthic foraminifers (Saidova, 1961). The C_{org} content in these sediments does not exceed 2%; the highest concentrations (to 1.5–2%) are confined to some areas at the base of the continental slope of Sakhalin and Kamchatka in the depth interval of 150–3360 m (Bezrukov, 1960). Sediments with the high C_{org} content are also marked by the elevated hydrotroillite content and H_2S contamination (Ostroumov, 1957).

RESULTS AND DISCUSSION

Benthic foraminifers are present in all the samples studied. Their total abundance, taxonomic composition and structure of assemblages show significant variations throughout the section (Figs. 3, 4). Based on the changes of these characteristics in the sediment sequence, three main benthic foraminiferal assemblages different in their structure can be distinguished (from the base upward).

The *Eponides weddellensis* assemblage (I) occurs in the interval of 330–260 cm. Sediments of this interval are represented by gray fine-grained mud with an admixture of sandy material and with the lowest $CaCO_3$ and C_{org} contents: commonly less than 2 (averaging 1.5) and 1.2%, respectively (Fig. 2). The total abundance of foraminiferal shells is also persistently low (100–150 specimens/g of dry sediment). The foraminiferal assemblage in this interval consists of 33 species, whose abundance ranges from 16 to 27 species per sample. The strongly dominant species is *Eponides weddellensis* (to 35–40%). A lower but significant role in this assemblage belongs to *Elphidium clavatum* (to

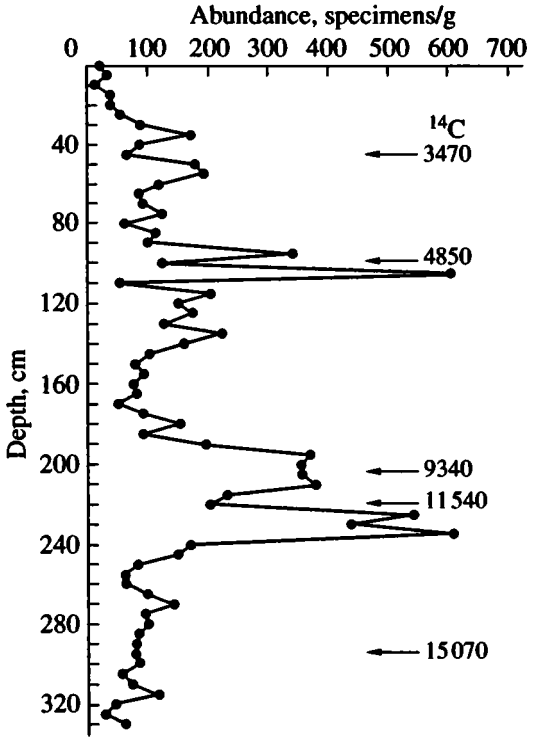

Fig. 3. Variations in the benthic foraminifer abundance (specimens/g of dry sediment) through the core V34-98. Arrows show the radiocarbon-dated levels in the core (years).

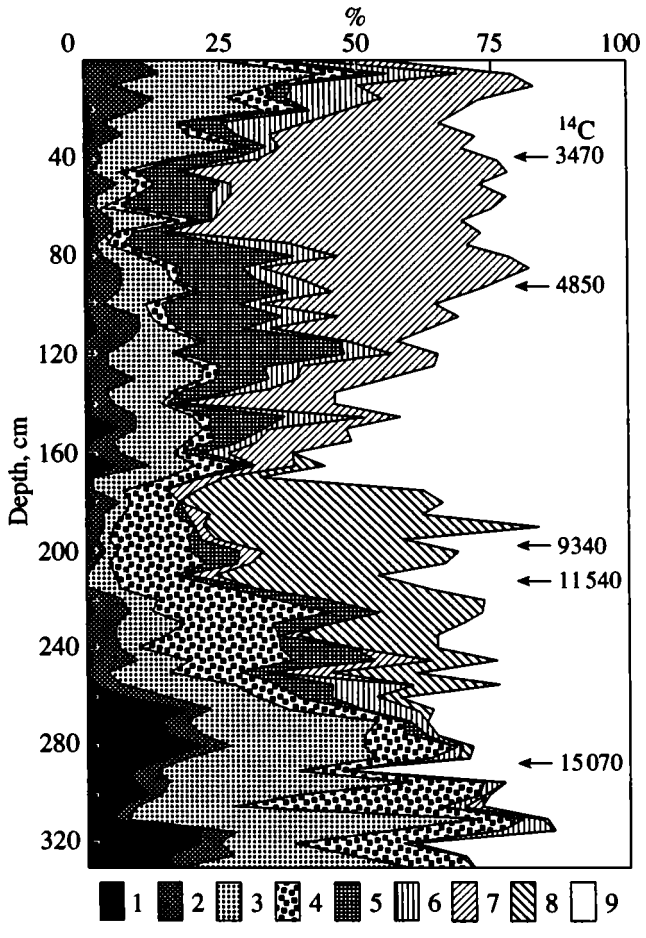

Fig. 4. Changes in the structure of benthic foraminiferal assemblages from the core V34-98: (1) *Elphidium clavatum*; (2) *Valvulineria ochotica*; (3) *Eponides weddellensis*; (4) *Uvigerina peregrina*; (5) *Cassidulina laevigata*; (6) *Cassidulina teretis*; (7) *Bolivina seminuda*; (8) *Brizalina spissa*; (9) other species. Arrows show the radiocarbon-dated levels in the core (years).

15–20%), *Uvigerina peregrina* (10–40%), *Nonion labradoricum* (5–15%), *Cassidulina teretis* (5–10%), and *Valvulineria ochotica* (about 5%). In total, the listed species usually compose more than half of the assemblage (sometimes to 90%).

The *Brizalina spissa* assemblage (II) is present in the interval of 260–180 cm. Sediments of this interval are composed of dark gray pelitic silty mud showing a greenish shadow and an insignificant sandy admixture. The $CaCO_3$ content sharply increases here as compared with underlying sediments and varies from 1 to 8%. There are two layers with an elevated $CaCO_3$ content in the lower (230–240 cm, to 7%) and upper parts (190–210 cm, to 8%) of the member separated by sediments with the $CaCO_3$ content of 3–4% (210–225 cm). The C_{org} content also substantially increases. This parameter is highest (to 1.5–1.7%) in the lower and upper layers of the member. The layer of 210–225 cm is marked by the C_{org} decrease to 1.2% (Fig. 2). Similar patterns are characteristic of the distribution of the total benthic foraminifer abundance in sediments of this interval, which varies from less than 100 to 600 specimens/g and more. The diversity of benthic foraminifers ranges from 20 to 26 species per sample. The lowest diversity is noted in the interval of 215–220 cm.

The taxonomic composition of benthic foraminifers also drastically changes in the considered interval. The assemblage is dominated by *Brizalina spissa*, which appears as single specimens in the uppermost part of the previous interval to make up 30–40% on average and to reach, sometimes, 80% of the total abundance in the sediments. *Cassidulina laevigata*, which was recorded only as an accessory species in two samples from the topmost part of the underlying member, represents a new subdominant element in this foraminiferal assemblage. In the considered interval, this species occurs persistently, sometimes amounting to 10% of the total abundance. An exception is the layer of 215–220 cm, where the species is missing. *Uvigerina peregrina* continues to play a significant role: its content averages about 10% and is sometimes as high as 30%. Common also are *Eponides weddellensis* (5–10%), the dominant species of the previous interval, and *Valvulineria ochotica* (5% and less). *Elphidium clavatum* and *Cassidulina teretis* almost completely disappear.

The *Bolivina seminuda* assemblage (III) occurs in the upper 180 cm. Sediments of the interval are represented by light green pelitic diatomaceous ooze with a $CaCO_3$ content of 2–4% in the lower (155–180 cm) and upper (0–30 cm) parts of the interval and of 5–6% in its middle part. Distinct against this background is a layer at the level of 95 cm, where the $CaCO_3$ content reaches 9%, the maximum value for the entire core. The C_{org} content gradually increases from 0.5% at the base to 1.4–1.7% in the middle part and to 2% in the upper part of the considered member (Fig. 2).

The benthic foraminifer abundance within the interval significantly varies from 100 to 600 specimens/g. The curve of variations in the total foraminiferal abundance coincides, with rare exceptions, with the curve of the $CaCO_3$ distribution: the minimal shell concentration (100 specimens/g and less) is recorded in the lower and upper parts of the member, whereas the relatively high abundance (to 150–200 specimens/g) is confined to its middle part. The layer of 95–105 cm is marked by a sharp increase in the foraminiferal abundance to 350–600 specimens/g. Similar strong variations are also characteristic of the foraminiferal diversity, which ranges from 20 to 31 species per sample.

Similarly to the underlying sediments, the considered interval is clearly distinguished by the change in the composition of the dominant species. The lower boundary of the interval is distinctly marked by the disappearance of *Brizalina spissa*, the dominant species of the previous assemblage, at the level of 175 cm, as well as by a drastic abundance decrease of *Uvigerina peregrina*, whose relative content commonly does not exceed

a few percent in higher sediments. *Bolivina seminuda*, which is present in underlying sediments as an accessory species only at some levels, becomes the dominant species in the considered interval. Already in its basal layers, the relative content of this species makes up about 20% and beginning from the layer of 140–150 cm, it represents 40–50% of the assemblage, decreasing again to 15% in the upper 20-cm-thick layer. The significant role in the assemblage belongs alternatively to *Cassidulina laevigata* and *C. teretis*. The basal layer (170–180 cm) encloses only *C. teretis*. Upsection, to the level of 20 cm, both species occur together, but the content of *C. laevigata* is substantially higher then that of *C. teretis* (20–25% versus 5–10%). The upper 20 cm of the core show only the presence of the latter whose share increases to 20%; i.e., it becomes prevailing. The upper part of the section (0–40 cm) is also characterized by an elevated (to 20–30%) proportion of *Eponides weddellensis*, whose content is usually less than 10–15% in the lower layers of the considered interval. Persistently occurring is *Valvulineria ochotica*, whose content commonly does not exceed 5% and increases to 13% only in the upper 20 cm of the core.

Thus, sediments of the studied core contain three benthic foraminiferal assemblages, which distinctly differ in the structure and successively replace each other in the section. Boundaries between these assemblages are located at levels of 180 cm (between assemblages II and III) and 260 cm (between assemblages I and II). According to the interpolation of radiocarbon dates, the age of these boundaries is estimated to be 12.5 and 8.5 ka, respectively. In other words, the established assemblages correspond to three main periods of the late Pleistocene–Holocene history of the basin development, or to three climatostratigraphic intervals: the terminal stage of the last glaciation (17–12.5 ka), the deglaciation period (12.5–8.5 ka) and the last interglacial or postdeglaciation epoch (last 8.5 ka).

The presented materials on changes in the total abundance, composition, and structure of foraminiferal assemblages elucidate main stages in the evolution of the hydrologic regime in the Sea of Okhotsk during the late Pleistocene and Holocene. Interpretation of ecological parameters of the established benthic foraminiferal assemblages is possible based on direct analogies with their modern communities, whose distribution at the bottom of the Sea of Okhotsk is well known, owing to works by Saidova (1961, 1994, 1997). According to her data, shelf assemblages influenced by the cold Okhotsk-Sea water mass are dominated by *Elphidium clavatum, E. orbiculare, Nonion labradoricum, Uvigerina peregrina*, and *Cassidulina smechovi*. The latter taxon represents a junior synonym of *C. teretis* described by Tappan (1951). In assemblages populating the shelf-break zone and continental slope influenced by warmer Pacific waters, the dominant species are represented by *Valvulineria ochotica, Uvigerina peregrina, U. auberiana, Eponides weddellensis, Cassidulina laevigata*, and *C. teretis*. According to Saidova (1961), the bottom areas, where sediments are enriched in an organic matter, are usually occupied by a community, which is characterized by the prevalence of *Uvigerina peregrina* and a high total foraminiferal abundance. Deep areas of the basin are inhabited by a community with *Eponides weddellensis* playing a significant role. Rises in the central part of the sea and areas of the continental slope of the Kamchatka, Sakhalin, and Kuril Islands are occupied by a community with the dominant species being *Cassidulina teretis*. The studied core was taken precisely in the area populated by the latter assemblage. Southward, near the Kruzenshtern Strait, the assemblage dominated by *Cassidulina laevigata* is distributed.

An analysis of benthic foraminifers reveals that the hydrological regime in the considered area, during the terminal stage of the last glaciation, was substantially different from the modern one. This resulted in the development of a peculiar foraminiferal assemblage, whose analogs are unknown in recent sediments of the Sea of Okhotsk (Fig. 4). *Eponides weddellensis*, which is the dominant species amounting to 45% of the assemblage from glacial interval sediments, persistently occurs at the correlative depth in the modern sea, but always plays a subordinate role. An exception is a foraminiferal assemblage that populates the bottom of the Kuril Basin filled with the deep water mass cooled to 1.8°C. Other characteristic species of this assemblage (*Uvigerina peregrina, Cassidulina teretis, Nonion labradoricum*) are widespread now, both in coastal areas and in the central part of the Sea of Okhotsk. The remarkable feature of these taxa is their tolerance for low temperatures. For instance, Saidova (1961) notes that *C. teretis* (*C. smechovi*, according to her determination) substantially increases within coastal areas of the northwestern Pacific (Kamchatka, Japan) influenced by cold currents. The increase in abundance of this species is also recorded in modern sediments of the continental shelf of the Greenland, North, and Barents seas, where the temperature of bottom waters is around zero (Mackensen and Hald, 1988), as well as in upper Pleistocene sediments of the North Atlantic deposited during cooling periods (Feyling-Hanssen *et al.*, 1983; Murray, 1984; Schnitker, 1984). Unlike the abovementioned species, *Elphidium clavatum*, whose share in glacial sediments amounts to 10–15%, is mainly distributed in modern sediments of the northern shelf of the Sea of Okhotsk, where they are influenced by the Okhotsk-Sea water mass of the temperature and salinity (Saidova, 1961). It also represents the secretory benthic foraminiferal species most widespread in the Arctic seas (Nagy, 1965; Feyling-Hanssen, 1976; Ostby and Nagy, 1982; Polyak, 1985; Lagoe, 1991).

Thus, the foraminiferal assemblage of the last glaciation was mainly represented by species that recently occurred as accessory elements in different communities from the deep water assemblages inhabiting the bottom of continental slopes and open sea areas to coastal ones. Nonetheless, despite the differences in

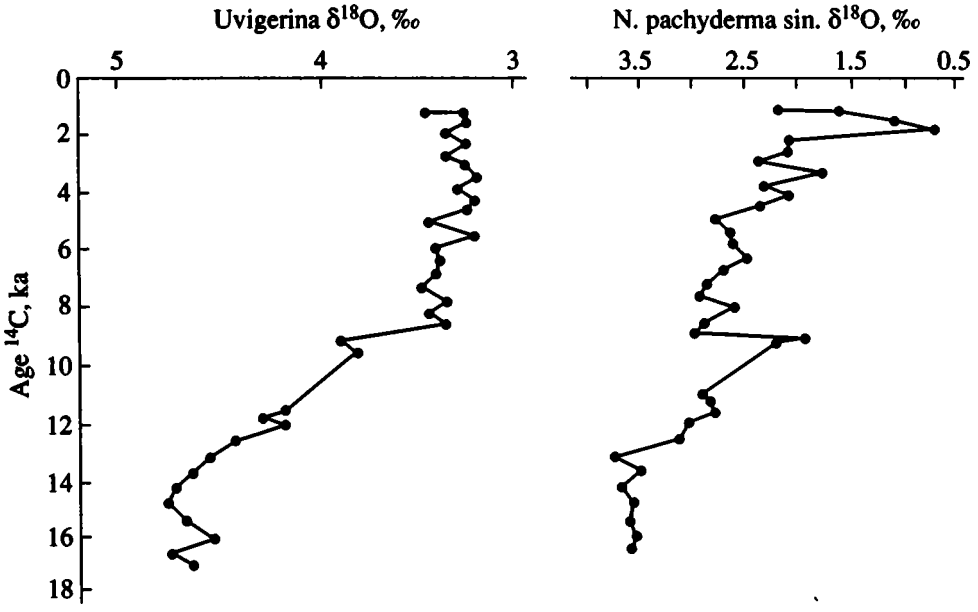

Fig. 5. $\delta^{18}O$ changes in *Uvigerina* spp. and *Neogloboquadrina pachyderma* sin. (Gorbarenko *et al.*, in press).

their inhabitancy depths, all of them have a feature in common—their tolerance to low and even negative temperatures of bottom waters. This allows the suggestion that in the period from 17 to 12.5 ka, i.e., at the terminal stage of the last glaciation, the Sea of Okhotsk was filled with waters close, in their temperature characteristics, to the modern Okhotsk-Sea water mass. Isotopic studies of *Uvigerina* species (*U. peregrina, U. senticosa, U. akitaensis*) confirm this assumption. Figure 5 demonstrates that this period was characterized by highest $\delta^{18}O$ values (Gorbarenko *et al.*, in press). The presence of abundant well-preserved shells of the typical Arctic species *Elphidium clavatum* probably indicates that seasonal ice covered also the southeastern part of the sea, because the good preservation of shells excludes their underwater redeposition from shallow depths. As evident from lithology (terrigenous silty mud), low $CaCO_3$ and C_{org} content (Fig. 2), as well as from the low foraminifer abundance, the productivity of surface waters during the entire period in question was probably insignificant. Deposition was characterized by low sedimentation rates (averaging 15 cm/1000 years) and occurred mainly under the influence of the terrigenous influx (Fig. 6).

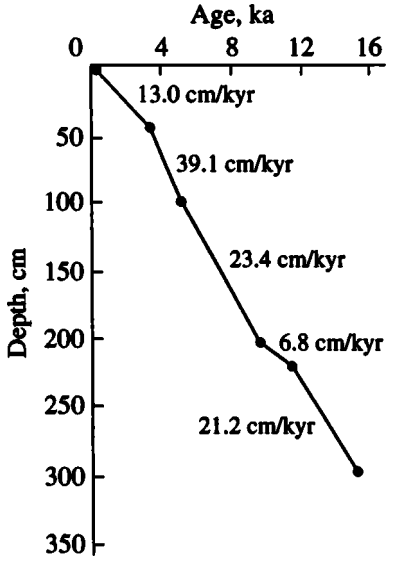

Fig. 6. Variations of sedimentation rate during the late Pleistocene–Holocene in the core V34-98 sampling area, the southeastern Sea of Okhotsk.

The onset of deglaciation (about 12.5 ka) was marked by drastic changes in the hydrological regime of the Sea of Okhotsk, which is reflected in the benthic foraminifer composition: the latter unambiguously indicates a noticeable rise of bottom water temperature. This is first evident from a complete disappearance of the Arctic species *Elphidium clavatum* and from a sharp decrease in the share of *Eponides weddellensis*, the typical species of modern deep-water basins of the Sea of Okhotsk. The leading role in the foraminiferal assemblage of the deglaciation period passed to *Brizalina spissa*, which occurs only as an accessory element in modern sediments of the considered sea basin. In the modern oceans, this species is abundant in high-productive areas (areas of hydrological fronts or zones of coastal upwellings). In the Pacific, rich populations of this species are recorded in the subarctic front area near Hokkaido Island and near the northern extremity of Honshu Island at depths of 1000–1400 m (Saidova, 1961). High concentrations of *B. spissa* are characteris-

tic of the coastal upwelling zone along North, Central, and South America, where sediments show very high C_{org} contents, whereas bottom waters are impoverished in dissolved oxygen (less than 0.5 ml/l) (Bandy *et al.*, 1960; Smith, 1963, 1964; Phleger and Soutar, 1973; Khusid, 1977, 1979). *Uvigerina peregrina* from recent sediments of the Sea of Okhotsk is abundant in coastal areas enriched in organic matter (Saidova, 1961). Similarly to *Brizalina spissa*, the most abundant populations of the above species are characteristic of high-productive areas in the oceans, where the species usually dwell on bottom areas with the elevated organic matter content (Phleger and Soutar, 1973; Khusid, 1979; Miller and Lohman, 1982; Mackensen *et al.*, 1985, 1990; Loubere, 1991).

The warming of the bottom waters is also evident from the appearance of relatively thermophilic *Cassidulina laevigata* already at the beginning of deglaciation (at the level of 260 cm); higher, this fossil becomes one of the dominant species. In the modern-day Sea of Okhotsk, the species dwells in bottom areas 650–1500 m deep and influenced by warm Pacific waters with temperatures above 2°C. In other areas of the World Ocean, it also prefers elevated bottom water temperatures. For instance, in the North and Norwegian seas, *C. laevigata* occurs only in areas affected by warm Atlantic waters (Mackensen and Hald, 1988). An important role in the foraminiferal assemblage belongs also to *Valvulineria ochotica*, the species, which is distributed now in the central part of the sea at a depth of 700 to 1700 m influenced by Pacific waters.

The above-mentioned data indicate that the glacial period termination was marked in the Sea of Okhotsk by the warming and intensified water mass circulation. As a result, Pacific waters that entered via the deep-water Kruzenshtern Strait began to fill the intermediate depth levels of the sea. This was accompanied by colonization of the sea bottom by a foraminiferal assemblage, whose important component was *Cassidulina laevigata*. Warming of bottom waters is also distinctly registered by the significantly lighter oxygen isotopic composition, which became evident 12–12.5 ka ago (Fig. 5). This event was likely associated with the surface water influx, which resulted in the warming, elevated bioproductivity, and increased supply of organic matter to bottom sediments (Fig. 2). This is particularly reflected in an increase of the total abundance of benthic foraminifers and in a strong preponderance of *Brizalina spissa* and *Uvigerina peregrina*. As was mentioned with respect to modern oceans, these species are widely distributed in the high-productive areas. The increased past productivity of surface waters is also evident from the appearance of a noticeable amount of siliceous microfossils in sediments, in which they later acquired the sediment-forming role. The rise of the surface water temperature in the Sea of Okhotsk is well reflected in the negative $\delta^{18}O$ excursion (by about 1‰) in the isotope curve compiled for core V34-98 on the basis of changes in the oxygen isotope composition of planktonic species *Neogloboquadrina pachyderma* sin. (Gorbarenko *et al.*, in press). The $\delta^{18}O$ decrease in planktonic foraminiferal shells is also established in core V34-90 from the Akademii Nauk Rise (Gorbarenko *et al.*, 1998; Chekhovskaya and Basov, 1999). A similar event is also registered in the North Pacific (Zahn *et al.*, 1991; Keigwin *et al.*, 1992; Kennett, 1995; and others). Moreover, isotopic data suggest that the warming tendency during deglaciation was not linear, but followed the more complicated two-step pattern.

An analysis of benthic foraminifers reveals that the period under consideration was also marked in the study area by two distinct stages of bottom water warming, which are reflected in the highest abundance of these fossils, relatively high proportion of thermophilic *Cassidulina laevigata*, and in the maximum $CaCO_3$ content in sediments. The stages are separated in the section by an interval displaying a sharp (threefold) decrease of benthic foraminifer abundance and the disappearance of *C. laevigata* in the layer of 215–220 cm, although this form plays a significant role in the underlying and overlying sediments. Simultaneously, the $CaCO_3$ content substantially decreases in this interval, probably corresponding to a brief cooling event, which coincides with the Younger Dryas. The climatic event in question is readily recognized by isotopic studies in many areas of the northern hemisphere as a whole and in the North Pacific region, including the Sea of Okhotsk, in particular, although the event dates are slightly variable (Keigwin, 1987; Zahn *et al.*, 1991; Keigwin *et al.*, 1992; Kennett, 1995; Kennett and Ingram, 1995; Kennett and Venz, 1995; Gorbarenko *et al.*, 1998; Chekhovskaya and Basov, 1999). Recently, it has also been established in New Zealand (Broecker, 1997).

In curves demonstrating changes in the oxygen isotopic composition for planktonic and benthic foraminifers from Core V34-98, the above event is probably reflected in the relatively low-amplitude positive $\delta^{18}O$ excursion (by about 0.2‰) approximately 11.5–11.6 ka ago, which corresponds to the level of 220 cm in the section (Gorbarenko *et al.*, in press; see Fig. 5). It should be noted that in Core V34-90 located southwestward at the Akademii Nauk Rise, this excursion is manifested significantly better, by 0.5‰ (Gorbarenko *et al.*, 1998; Chekhovskaya and Basov, 1999). This is probably related to the fact that Core V34-98 is located in the area near the Kruzenshtern Strait directly influenced by entering Pacific waters. Precise dating of the event in the studied core is difficult. Its interpolated age is about 11.5 ka; however, this date is most likely overestimated because of the hiatus or strongly condensed sedimentation near this level. The hiatus in the interval of 204–219 cm is also noted by (Gorbarenko *et al.*, in press) and can be inferred from the curve of sedimentation rate. In Fig. 6, it is seen that the rate sharply decreases in this interval. The weak manifestation of the event in the isotopic curve is probably an effect of the hiatus. It should be noted that the hiatus itself can indicate an intensification of bottom currents due to sharp cooling.

The deglaciation period is marked by a general decrease in sedimentation rates to 23 cm/kyr that may be also indicative of intensified erosion on surrounding land and an increase in the terrigenous influx to the basin.

At the beginning of postdeglaciation (the commencement of the current interglacial period about 8.5 ka ago), the bottom of the Sea of Okhotsk was populated by the foraminiferal assemblage, whose composition indicates a further transformation of the hydrological regime in the basin (Fig. 4). Similarly to the glaciation-deglaciation boundary time, changes in the foraminiferal community at the onset of the postdeglaciation period were sharp. The basal layer of interglacial sediments (upper 180 cm) is already devoid of *Brizalina spissa*, which represented almost a half of the previous foraminiferal assemblage. Simultaneously, the abundance of *Uvigerina peregrina* substantially decreases. The benthic foraminiferal assemblage dominated by *Bolivina seminuda, Cassidulina laevigata, C. teretis, Eponides weddellensis*, and *Valvulineria ochotica* suggests that warm Pacific waters continued to persistently enter the Sea of Okhotsk. The growth of bottom water temperatures at that time is also confirmed by a significant negative $\delta^{18}O$ excursion (Fig. 5).

The bottom hydrodynamics was apparently less active than during the preceding period, as is evident from the sediment lithology represented by fine-grained pelitic mud. Judging from the increased role of planktonic siliceous microfossils (diatoms and, to a lesser degree, radiolarians), the surface water productivity of the basin became even higher, as compared with the deglaciation period, and resulted in the increased (to 1.5–2%) C_{org} content in sediments (Fig. 2). A combination of high bioproductivity, responsible for a very high organic matter influx, and a relatively calm hydrodynamic regime in the bottom water layer favored specific conditions, under which only some species had an advantageous position. *Bolivina seminuda* represents precisely such a species. In modern oceans, this species forms mass accumulations only in the very high-productive areas of coastal upwelling, which are characterized by the high C_{org} content in sediments and by deficiency of the dissolved oxygen in bottom waters that is typical, for instance, of the Californian and Peruvian coasts of the Pacific (Uchio, 1960; Phleger and Soutar, 1973; Khusid, 1979). In the modern Sea of Okhotsk, *B. seminuda* occurs as single specimens in the depth interval of 850–1700 m within areas influenced by Pacific waters (Saidova, 1961). A strong preponderance of this species in the foraminiferal assemblage (to 40–50%), particularly in the time interval of about 2–5 ka ago, allows an assumption that the relatively warm bottom waters were characterized at that time by the lowered content of dissolved oxygen, probably because the latter has been consumed for oxidation of large volumes of organic matter in bottom sediments, which accumulated in response to the high plankton productivity in surface waters.

Temperature characteristics of Pacific waters in the studied area during the entire postglacial period were probably also variable. Judging from the absence of relatively thermophilic *Cassidulina laevigata* among benthic foraminifers, the temperature values at the deglaciation–postdeglaciation boundary were minimal. This is also evident from the decreased total foraminiferal abundance. In the middle Holocene (approximately between 6 and 3.5 ka), the bottom water temperatures were probably at the maximum in the considered time interval (the last 17 kyr), as is evident from the growth of the total foraminiferal abundance against the background of sharply enhanced sedimentation rates (to 39 cm/kyr), and also from the substantial increase of the *C. laevigata* share in the assemblage. In the late Holocene (the last 3–3.5 kyr), sedimentation rates fell again and decreased the foraminiferal abundance. The event was responsible for replacement of relatively thermophilic *C. laevigata* by cold-resistant *C. teretis* and for a wide development of deep-water *Eponides weddellensis*. This assumption is supported by changes of the oxygen isotope composition measured in *Uvigerina* shells (Fig. 5). The late Holocene was marked by the decline in the sedimentation rate to 13 cm/kyr. As is evident from the sediment lithology (diatomaceous ooze), this was a response to the decreased terrigenous influx, which can indicate a further lowering of hydrodynamic activity in the bottom layer.

CONCLUSION

Our analysis of variations in the total abundance, taxonomic composition and structure of benthic foraminiferal assemblages in the upper Pleistocene–Holocene sediments obtained from the continental slope of the Kuril Islands in the southeastern Sea of Okhotsk (Core V34-98) allows the following conclusions to be drawn:

(1) During the late Pleistocene–Holocene, the floor of the Sea of Okhotsk was successively populated by three benthic foraminiferal assemblages, which sharply differ from each other by dominant species and are termed as the *Eponides weddellensis* (late Pleistocene), *Brizalina spissa* (latest Pleistocene–earliest Holocene) and *Bolivina seminuda* (Holocene) assemblages. Everyone of the assemblages corresponds to a particular stage in the climatic and hydrological evolution of the basin.

(2) Based on ecological characteristics of benthic foraminiferal assemblages and radiocarbon dating, it is established that the distinguished assemblages correspond to the last glaciation (17–12.5 ka), deglaciation (12.5–8.5 ka) and postdeglaciation, or the current interglacial period (the last 8.5 kyr).

(3) The results of foraminiferal analysis indicate that during the glacial period bottom waters of the Sea

of Okhotsk were characterized by low, probably negaive temperatures. Under these conditions, only species that now dwell in areas influenced by the Okhotsk-Sea water mass and also in the Arctic sea basins could survive. The sea surface, including its southeastern area, was probably covered by ice during the greater part of the year, and this caused low bioproductivity in the surface water layer and, as a result, an insignificant influx of organic matter and remnants of planktonic siliceous and calcareous microorganisms into sediments. Sedimentation, which was characterized by relatively low rates, was mainly controlled by the runoff from the surrounding slightly elevated land.

(4) The glaciation–deglaciation boundary (12.5 ka ago) was marked by a general climatic warming and changes in the bottom (and surface) circulation. At this time, the Sea of Okhotsk basin began to be filled with warm Pacific waters that entered via the Kruzenshtern Strait in the northern part of the Kuril island chain and were characterized by the temperature values of 2°C and higher. These processes resulted in a significant growth of bioproductivity in surface waters and in the increased organic matter influx to bottom sediments. They are also reflected in corresponding increase in diversity and abundance of benthic foraminifers, whose assemblages show the dominant role of more thermophilic, as compared with the glacial period, species characteristic of high-productive areas of the modern oceans. An enhanced erosion on surrounding land due to warming and deglaciation, together with the corresponding increase of the terrigenous influx and growing productivity of calcareous and siliceous microplankton, resulted in the sedimentation rate increase and elevated $CaCO_3$ and SiO_2 contents in sediments.

(5) Similarly to other areas of the Sea of Okhotsk and northern Pacific, where the two-step mode of deglaciation is evidenced by isotopic data, two warm phases separated by a brief cooling, which is approximately synchronous with the well-known worldwide Younger Dryas event, are established in the studied area based on benthic foraminifers.

(6) During the recent interglacial, or postdeglaciation period (last 8.5 kyr), the hydrodynamic activity in the bottom water layer was probably lower relative to that of the previous period, as it follows from the pelitic type of sediments and from peculiarities of benthic foraminiferal assemblage. The high microplankton productivity in surface waters provided a significant influx of biogenic (carbonate and siliceous) material and organic matter. The relatively low hydrodynamics in the bottom water layer and elevated nutrient content in sediments created the peculiar conditions advantageous only to some species. One of such species was *Bolivina seminuda*, whose mass accumulations in modern oceans occur only in the very high-productive areas of coastal upwelling, where environments are characterized by the high C_{org} content in sediments and by a deficiency of the dissolved oxygen in bottom waters. The near-bottom temperature in the considered area also experienced variations, although the influence of warm Pacific waters was continuous. As evidenced by distribution of the relatively thermophilic *Cassidulina laevigata*, the highest temperature values were characteristic of the period approximately between 6 and 3.5 ka.

(7) Changes in composition of benthic foraminiferal assemblages allow the assumption that the drastic hydrological transformations in bottom waters were short-term events that lasted several centuries during the glaciation–deglaciation and deglaciation–postdeglaciation boundary epochs.

(8) Climatic and hydrological changes in the Sea of Okhotsk that are revealed using the results of the qualitative and quantitative analysis of benthic foraminifers is consistent with data on the oxygen isotope composition in the foraminiferal shells.

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Stages in Evolution and Biogeography of Permian Ammonoids

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Abstract—The generic diversity of Permian ammonoids is analyzed, and four evolutionary stages of the fossils are distinguished. The diversity dynamics is elucidated for five principal distribution areas: the Paleotethyan, Uralian, American, Arctic, and Australian provinces. Certain trends of ammonoid biogeographic distribution are characterized for principal stages of their evolution.

Key words: *biogeography, ammonoids, evolutionary stages, Permian*

INTRODUCTION

Ammonoids represent the relatively well-studied group of fossils of high stratigraphic potential for subdivision and correlation of Upper Paleozoic deposits, Permian in particular, but despite this, the biogeographic distribution of these pelagic (or benthonic–pelagic) organisms has so far been inadequately studied. The problem has been analyzed only for certain time intervals of the Permian System in separate areas (Zhou, 1986; Spinosa *et al.*, 1991; Bledinger *et al.*, 1992; and others). The circumstantial paper by Nassichuk (1995) is devoted to detailed characteristics of Permian ammonoid fauna from the Arctic regions of the globe. A group of Chinese and American scientists (Zhou *et al.*, 1996) have recently analyzed the evolutionary changes of ammonoids, disregarding their geographic differentiation. Unfortunately, they excluded the Kungurian Age from consideration and thus presented the less precise and unfinished general evolutionary patterns of ammonoids. It is also difficult to agree with the taxonomy and stratigraphic ranges of many Permian ammonoid taxa accepted in their work, as these issues are considered at the knowledge level of the 1960s.

This study is aimed, first of all, to generalize the available data on geographic ranges of Permian ammonoids, although this problem cannot be fully solved at present. Natural zones of the Permian epoch were much more differentiated as compared to those of the Carboniferous. The Permian evolution history of ammonoids was quite variable, full of significant events, some of which were dramatic (the appearance of the new ammonoid order Ceratitida at the Early–Late Permian boundary and the almost complete extinction of the whole subclass near to the Permian–Triassic boundary). Biogeographic connections remained unstable during the Permian and changed from one stage to another in the course of ammonoid evolution.

EVOLUTIONARY STAGES OF PERMIAN AMMONOIDS

At the commencement of the Permian, the most widespread were representatives of ammonoid orders Goniatitida and Prolecanitida, consisting of mostly families and genera inherited from the Carboniferous. The very beginning of the Permian was marked by the appearance of four new families: Paragastrioceratidae (genus *Svetlanoceras*), Metalegoceratidae (genus *Juresanites*), Perrinitidae (genus *Properrinites*), and Popanoceratidae (genus *Protopopanoceras*). In addition, some new genera appeared in two families that had existed before: genus *Kargalites* amid Marathonitidae, and genera *Prostacheoceras* and *Tabantalites* amid Vidrioceratidae.

In analyzing changes in the generic diversity, morphological evolution, and events of mass extinction, I distinguished four principal stages of the Permian ammonoid history (Fig. 1). Four evolutionary stages have been also recognized in the history of other marine invertebrates, such as fusulinids, conodonts, and brachiopods (Leven *et al.*, 1996).

The first "pro-Permian" stage spans the interval of the Asselian and Sakmarian ages, together with a part of the Artinskian Age. In general, the ammonoid genera of this period were rather similar in taxonomic composition and morphology to the Carboniferous ancestors. Their evolution was characterized by a considerable growth of taxa (more than twice) in both aforementioned orders. The total number of genera increased from 17 to 37: from 12 to 25 in the order Goniatitida (13 new genera) and from 5 to 12 in the order Prolecanitida (7 new genera). The greater taxonomic diversity was accompanied by a certain sophistication of morphology as compared to that of the Carboniferous goniatites, who mostly had the eight-lobe septa (in the suture of goniatites from families Marathonitidae, Vidrioceratidae, and Shumarditidae, the amount of lobes becomes greater and these elements show the second-

ary segmentation like in representatives of newly appeared families). The development of a more sophisticated hydrostatic apparatus allowed ammonoids to better adapt to the diverse habitat environments. The significant extinction at the end of the first stage primarily hit genera originated in the Carboniferous.

The second "eo-Permian" stage spanning the Artinskian (partly) and Kungurian ages is characterized by the evolution of typical Permian goniatites and prolecanites. New genera, which appeared during this stage, represent almost 60% relative to the total number of generic taxa, whereas genera inherited from the Carboniferous exemplify less than 10%. The taxonomic diversity of both ammonoid orders grew further and was at its peak in the Kungurian Age. The total number of genera (Fig. 1) was 55 in the Artinskian Age (36 of goniatites and 19 of prolecanites) and 71 in the Kungurian time (57 of goniatites and 14 of prolecanites). The diversification progressed together with the significant morphological sophistication (among goniatites, there appeared the genus *Perrinites* with the "ammonite" suture, first multilobate *Stacheoceras* forms, and genera *Popanoceras* and *Thalassoceras* displaying the extremely segmented lobes; newcomers of prolecanites were genera *Bamyaniceras* and *Sicanites*, and also complex *Medlicottia* and the first *Eumedlicottia* forms).

The third "meso-Permian" stage corresponds to the interval of the Roadian and Capitanian ages (I consider the latter in a broad sense as including the Amarassian Stage). The main event of this period was the appearance of the first representatives from the genus *Paraceltites* of the new ammonoid order Ceratitida; these forms were initially primitive and scarce. Goniatites and prolecanites of that time showed the most complicated morphology (family Cyclolobidae of goniatites with genera *Waagenoceras, Timorites*, and *Cyclolobus*, and also genera *Eumedlicottia, Neogeoceras*, and some others of prolecanites), although there was a distinct tendency in the diversity decrease in both groups of fossils. In general, 50 Roadian genera (40 of goniatites, 9 of prolecanites, and one of ceratites), 50 Wordian genera (39 of goniatites, 8 of prolecanites, and 3 of ceratites), and only 24 Capitanian genera (16 of goniatites, 5 of prolecanites, and 3 of ceratites) have been identified.

The fourth "meta-Permian" stage spans the two last ages of the Permian: Dzhulfian and Dorashamian (or Wuchapingian and Changsingian, according to the scale suggested by the International Subcommission on Permian Stratigraphy). The wide distribution and abundance of diverse ceratites in the eastern Paleotethys, together with the parallel decline of goniatites and prolecanites, were the most characteristic features of the stage. Ammonoids of two latter groups became extinct almost entirely at the Permian–Triassic boundary. The abundance parameters for the fourth stage are as follows: 47 Dzhulfian genera (11 of goniatites, 4 of prolecanites, and 32 of ceratites) and 29 Dorashamian genera (4 of goniatites, one of prolecanites, and 24 of ceratites).

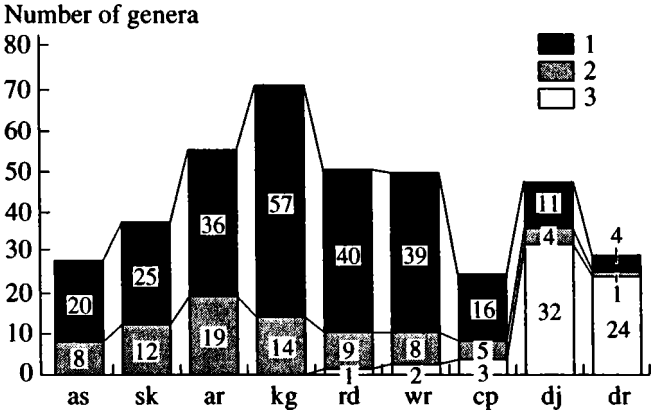

Fig. 1. Changes in generic diversity of Permian ammonoids across stages of the Permian System: (1) goniatites; (2) prolecanites; (3) ceratites. Stages: (as) Asselian; (sk) Sakmarian; (ar) Artinskian; (kg) Kungurian; (rd) Roadian; (wr) Wordian; (cp) Capitanian; (dj) Dzhulfian; (dr) Dorashamian.

Only two ceratitid genera crossed the Permian–Triassic boundary, but their representatives violently evolved and experienced extreme diversification during the Triassic. The last goniatites are known from the Dorashamian (Changsingian) deposits, and only one prolecanitid taxon (genus *Episageceras*) was able to survive this stage, though it disappeared soon thereafter and left no descendants.

BIOGEOGRAPHIC ANALYSIS

Experts in ammonoids used to consider two large distribution areas of their Permian representatives: Boreal (high latitudes with temperate climate) and Tethyan (warm-water basins of low latitudes) provinces. This subdivision seems to be quite conventional, and Nassichuk (1995) is obviously right in stating that the general term "Boreal province" denotes territories adjacent to the Arctic ocean, although there are exceptions (Nevada, North California, northern China). The Tethyan province is not well defined in the paleogeographic sense. Its separate areas recurrently changed their position in the course of geological history, so it is often difficult to understand where sediments bearing this and that fauna have been accumulated. The distribution of Paragastrioceratidae forms in "Boreal" regions and Perrinitidae species in "Tethyan" areas is usually considered as one of basic criteria discriminating different habitat zones of ammonoids. This discrimination is not unambiguous, however, because it is established that representatives of these two families coexisted in several regions. Besides, the Urals are known as a center of diversification and maximum abundance of paragastrioceratids cannot be attributed

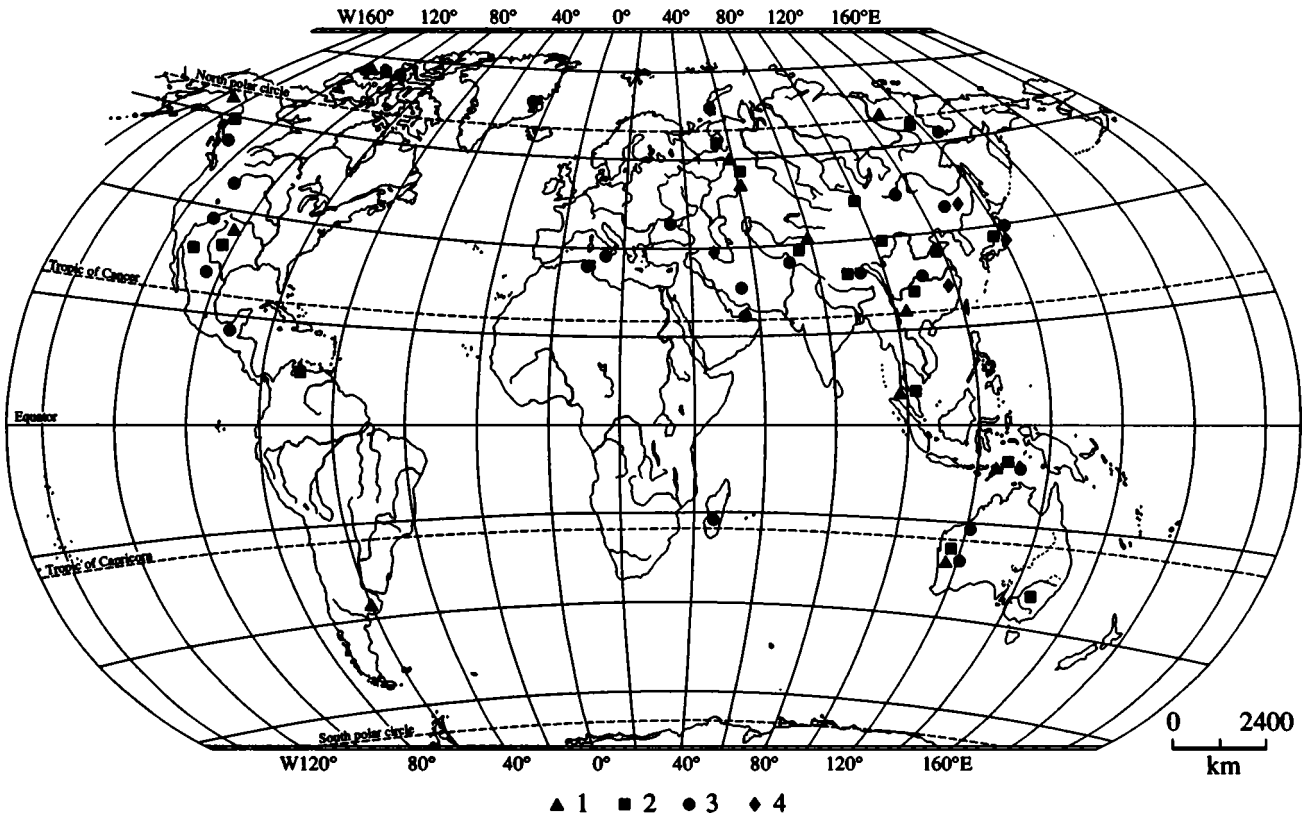
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Fig. 2. Main localities of Permian ammonoids: (1) first stage; (2) second stage; (3) third stage; (4) fourth stage.

unconditionally to the regions of the first type, although many researchers consider it as a part of the Boreal (Spinosa *et al.*, 1991; Nassichuk, 1995; and others) or Biarmian province (Ganelin and Kotlyar, 1984). At the very beginning of the Permian, the Uralian basin was freely connected with the Paleotethys, since the ammonoid species from genera *Synartinskia* and *Daixites*, both considered as typical endemics of the Early Permian fauna of the Urals, were also found in the Sakmarian deposits of the Darvaz (Leven *et al.*, 1992). Moreover, the genus *Cardiella*, widespread in basins of the Paleotethys, is known from the Artinskian sediments of the Urals. In the second half of the Early Permian, the connection between the Uralian basin and northern Paleotethys was destroyed. Data pointing to any connection between the Uralian and Boreal basins in the earliest Permian are unknown. Basins with fauna of the Boreal type were formed in the north of the Urals only at the end of the Early Permian. As for the American basins, the general opinion is that they were outside of the Tethyan province and populated by diverse ammonoid communities, the abundant perrinitids inclusive (Leonova, 1996). One may regard the differentiation of "Boreal" and "Tethyan" ammonoid faunas as dependent on such factors as the water temperature. Investigation of recent squids (Zuev *et al.*, 1996) showed, however, that these cephalopods dwell in water, where the temperature is not lower than +19°C, and the optimal range for them is +22...+24°C. Judging from these data, the problem to divide cephalopods into "thermophilic" and "cryophilic" groups appears to be difficult. It is likely that we may speculate about the issue only in terms of "more or less thermophilic" taxa. Thus, the current subdivision of ammonoids into Boreal and Tethyan forms does not correspond to the present knowledge of their biogeography. In order to elucidate the possible biochore system, I qualitatively analyzed the generic and familial composition, outlined the diversification centers, and detected quantitative changes in ammonoid communities. This and other data now available allowed me to recognize five biochores, whose ammonoid assemblages are representative, have distinctive taxonomic compositions, and show specific degree of endemism at the generic level (the table). The known localities of ammonoids characterizing different evolutionary stages are shown in the geographical map (Fig. 2) and in a series of paleogeographic schemes for the Sakmarian, Artinskian, Kazanian (Wordian), and Tatarian (Dorashamian) ages, which are based on the topography reconstructed by Ziegler *et al.* (1996) from the University of Chicago.

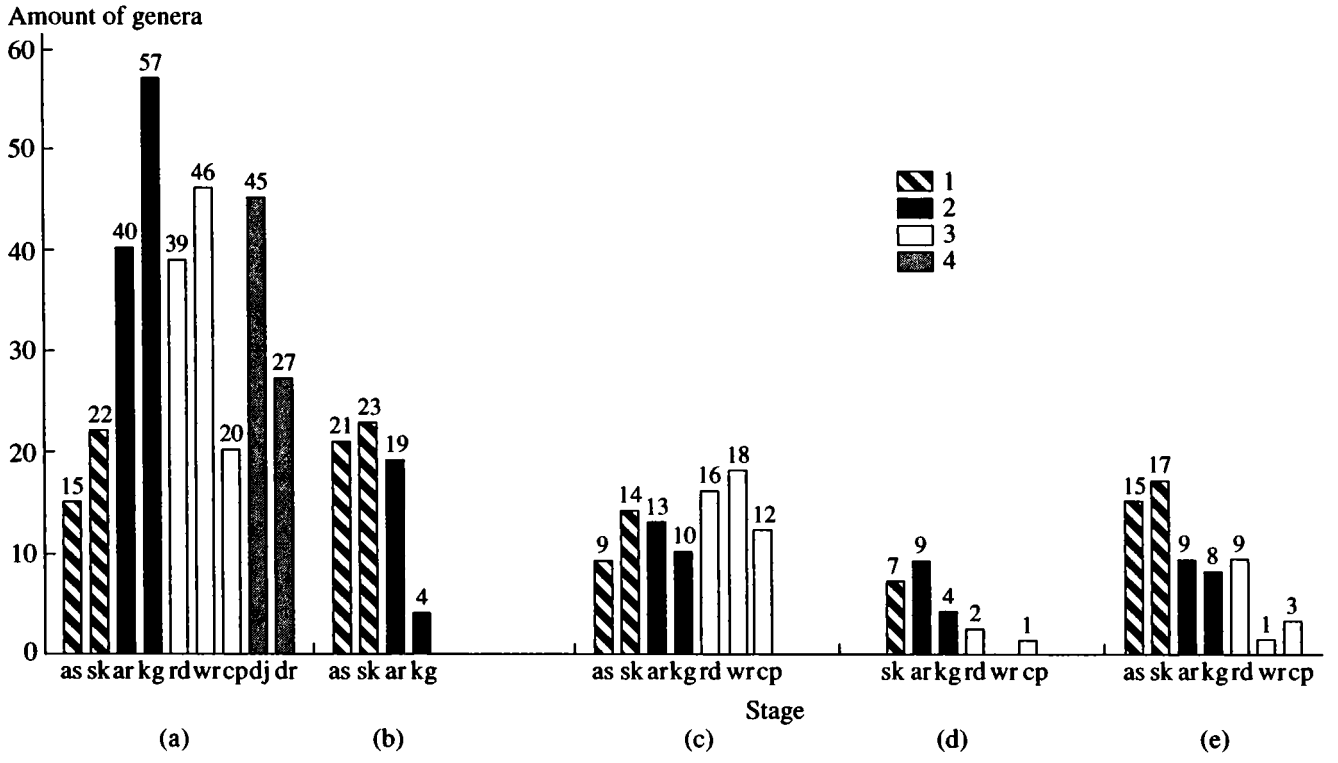


Fig. 3. Stage-related generic diversity of Permian ammonoids in various biogeographic areas: (1) first stage; (2) second stage; (3) third stage; (4) fourth stage; (a) Paleotethyan biochore; (b) Uralian biochore; (c) American biochore; (d) Arctic biochore; (e) Australian biochore. Stages: (as) Asselian; (sk) Sakmarian; (ar) Artinskian; (kg) Kungurian; (rd) Roadian; (wr) Wordian; (cp) Capitanian; (dj) Dzhulfian; (dr) Dorashamian (numbers above columns denote amount of genera).

PALEOTETHYAN BIOCHORE

The Paleotethys includes the ammonoid localities in Sicily, Crimea, Tunisia, Oman, Iraq (Kurdistan), Transcaucasia, Iran, Salt Range, Kashmir, Afghanistan, Pamirs, Darvaz, Tibet, China, Primor'e, Japan, Thailand, Malaysia, Timor, and Madagascar (Fig. 2). Most of ammonoid occurrences known at present are located in these regions, and they yielded the greatest number of Permian genera. The dynamics of generic diversity (Fig. 3a) is almost similar to that shown in the global scheme (Fig. 1).

At the first stage of the earliest Permian time (Fig. 4), 29 ammonoid genera known from the Paleotethys represented 45% of their total number, and only four of them were endemic (*Vanartinskia, Miklukhoceras, Parapronorites*, and *Martoceras*). Other genera were cosmopolitan, because this ocean, as mentioned above, was connected with the Uralian basins and also might have the fauna interchange with the American basins (indicative genera *Emilites* and *Properrinites*). According to available data, the Pamirs and Timor basins were the main diversification centers in the vast area of the Early Permian Paleotethys.

At the end of the Early Permian (second stage), more than 60 genera (or 80% of generic taxa of that time) inhabited the Paleotethys (Figs. 3a, 5). Nearly half of them, namely 28 genera were endemic. The late Artinskian–Kungurian ammonoid assemblages are the most representative as they include over 50 genera and 80 species of ammonoids. These were typical open-sea communities with abundant Medlicottiidae, Pronoritidae, Adrianitidae, and Perrinitidae. At that time, the basin of Pamirs was the largest diversification center, as it follows from the distribution analysis of medlicottiids (subfamily Sicanitinae, genera *Propinacoceras* and *Bamyaniceras* from subfamily Propinacoceratinae, and genus *Miklukhoceras*) and goniatites (family Perrinitidae; genera *Crimites, Veruzhites, Pseudoemilites*, and *Sosiocrimites* from family Adrianitidae; and also genera *Cardiella, Kargalites, Almites, Suakites*, and some others from the Marathonitidae family). Apparently, the geographic position of the Pamirs, between the eastern and western areas of the Paleotethys, was favorable, together with the free seaways for fauna interchange, for the origin and radiation of abundant new forms. The Timor and South China basins represented additional diversification centers inhabited by diverse and abundant ammonoid communities, where endemics represented a significant share of the population. Analysis of generic and specific composition of Artinskian and Kungurian ammonoids from the Pamirs, Darvaz, Afghanistan, southeastern China, Thailand, and Timor shows that there were persistent ways of fauna migration between these basins. Ammonoid assem-

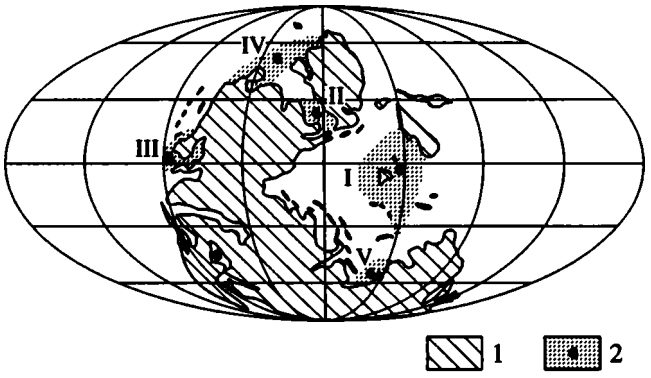


Fig. 4. Distribution areas of ammonoids at the beginning of the Early Permian (after Ziegler *et al.*, 1996, with additions): (1) land; (2) distribution areas of ammonoids; (I) Paleotethyan biochore; (II) Uralian biochore; (III) American biochore; (IV) Arctic biochore; (V) Australian biochore.

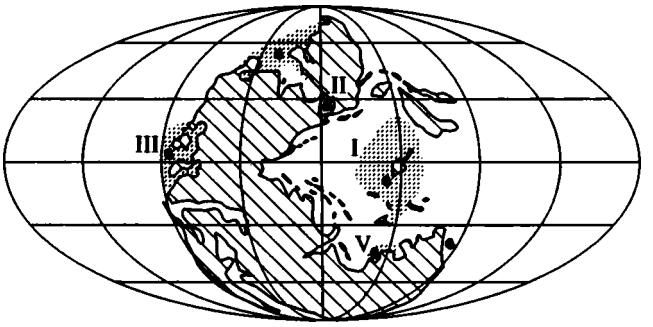


Fig. 5. Distribution areas of ammonoids at the end of the Early Permian (after Ziegler *et al.*, 1996, with additions; symbols as in Fig. 4).

blages from all these regions are rather similar in their taxonomic aspect, but they all show a high percentage of endemic forms. According to data on ammonoid distribution, basins of the Pamirs, Darvaz, and Afghanistan were situated near the South China block or between this structure and the Kimmerian epicontinent, but not as far away to the south and west from the southeastern China as is shown in palinspastic maps by McKerrow and Scotese (1990) or by Zonenshain *et al.* (1987). At any rate, there should be stable connections between the habitat areas of identical or close ammonoid communities. Investigation of present nautilus movement and distribution (Saunders and Spinosa, 1979) shows that the distribution areas of cephalopod species are not as great as was assumed in the concept of their free migration over the ocean. As it follows from observations, these areas have certain limitations and are not more than several hundreds kilometers across.

At the third stage of the early Late Permian (Fig. 6), the basins under consideration hosted 78–96% of all ammonoid genera, and 60–70% of this amount were represented by endemic taxa. At present, most representative distribution areas are known for the Wordian ammonoid faunas. Especially interesting are ammonoid assemblages from Sicily and the Timor. These two areas were situated in the extreme western (the former) and eastern (the latter) ends of a series of Wordian basins and likely represented the diversification centers from which many ammonoid forms migrated to meet one another. In this respect, the most spectacular examples are various Adrianitidae forms and close species from genera *Propinacoceras, Paraprononites*, and *Daraelites*. Between these centers, were located the basins of northern Iraq (Kurdistan), Oman, Tunisia, and Crimea, where abundant ammonoid communities consisted of identical or very similar species, and also the basins of Croatia and Chios Island (Greece), where their communities were less diverse. In the opinion of American experts (Bledinger *et al.*, 1992), there was also a seaway along the northern coast of Gondwana that connected the above areas and was favorable for the fauna interchange. To approve their idea, the cited researchers refer to data on coeval conodonts. The younger Amarassian ammonoids are known only from Timor Island, and this suggests that the formerly stable seaways were destroyed in the post-Wordian time.

Very diverse ammonoid assemblages of the initial Late Permian time are known from the basins of southeastern China. These assemblages are rich in endemic genera (over 50%), and the basins appear to have been partially isolated by certain barriers that provided the particular habitat environments for ammonoids. At the fourth stage of the terminal Permian time (Fig. 7), ammonoids survived in the Paleotethyan basins only (100%), as their remains are so far unknown from other localities. A series of these basins distinguished now in Transcaucasia, Iran, eastern and southeastern China, Far East, and Japan was populated by diverse and very peculiar ammonoid faunas. The Chinese scientist Zhou (1986) investigated in detail how the taxonomic composition of respective ammonoid communities depended on the type of their habitat basins. He demonstrated that the closed basins of the platform type and abnormal salinity accumulated in eastern and southeastern China the paralic coal-bearing deposits and were populated mostly by diverse ceratites, many of which were endemics and had the well-developed sculpture on shells peculiar in morphology. Several genera of goniatites (*Changhsingoceras, Pseudogastrioceras, Neoaganides*, and some others) and genus *Episageceras* of prolecanites were distinctly subordinate in these communities. Other ammonoid faunas very similar in taxonomic composition and general aspect are also known from the Upper Permian deposits of the Transcaucasia (Dzhulfa and Dorasham) and Iran (Abadeh). Sedimentation environments in all these

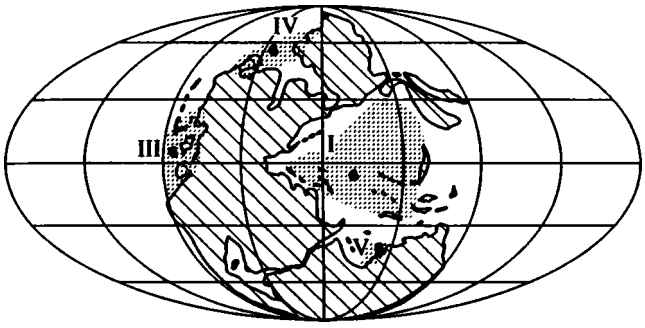


Fig. 6. Distribution areas of ammonoids at the beginning of the Late Permian (after Ziegler *et al.*, 1996, with additions; symbols as in Fig. 4).

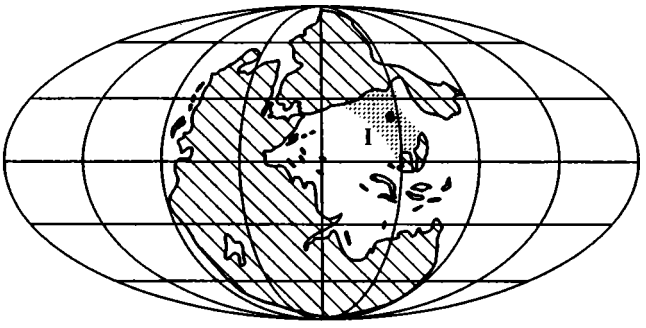


Fig. 7. Distribution areas of ammonoids at the beginning of the Late Permian (after Ziegler *et al.*, 1996, with additions; symbols as in Fig. 4).

areas were almost identical, and their shallow and closed epiplatform basins were interconnected through seaways suitable for migration of very specific ammonoids. The less representative ammonoid assemblages of the same type are described from the Kitakami area of Japan (Ehiro *et al.*, 1987, 1995) and from the Far East of Russia (Zakharov, 1994). It seems reasonable to take into account in all paleogeographic models that the regions of development of the Late Permian ammonoid fauna (Transcaucasia, Iran, South China, Japan, and the Far East) were located close to each other within a single biochore, being undivided by barriers impenetrable for these organisms.

URALIAN BIOCHORE

The Uralian biochore includes the ammonoid localities of central and southern Urals and of northern Kazakhstan (Fig. 2). In the basins of the Urals, ammonoids were especially diverse at the very beginning of the Permian (Asselian and Sakmarian ages), when they represented 64 and 59% of all coeval genera, respectively. Three families typical of the Permian (Paragastrioceratidae, Metalegoceratidae, and Popanoceratidae) and several new genera and species, which later became widespread in other Permian basins, originated in this biochore. Genera *Shikhanites, Neodimorphoceras, Protopopanoceras, Sakmarites*, and *Synuraloceras* are endemics and indicate that the Uralian biochore was the largest diversification center of the earliest Permian ammonoids. During the second stage, the relative diversity of ammonoids decreased, first to 34% (Artinskian Age) and then to 5.5% (Kungurian Age). The Artinskian Age was marked by a wide development of Paragastrioceratidae (genera *Paragastrioceras* and *Uraloceras*), which coexisted with diverse goniatites and prolecanites (endemic genera *Sakmarites, Aktubinskia*, and *Artioceras*). The Asselian and Sakmarian ammonoids are known mostly from southern areas of the Urals, whereas their Artinskian and Kungurian taxa are reported from more northern areas of the central Urals. Ammonoids postdating the Kungurian Age are unknown from the biochore in question (Fig. 3b).

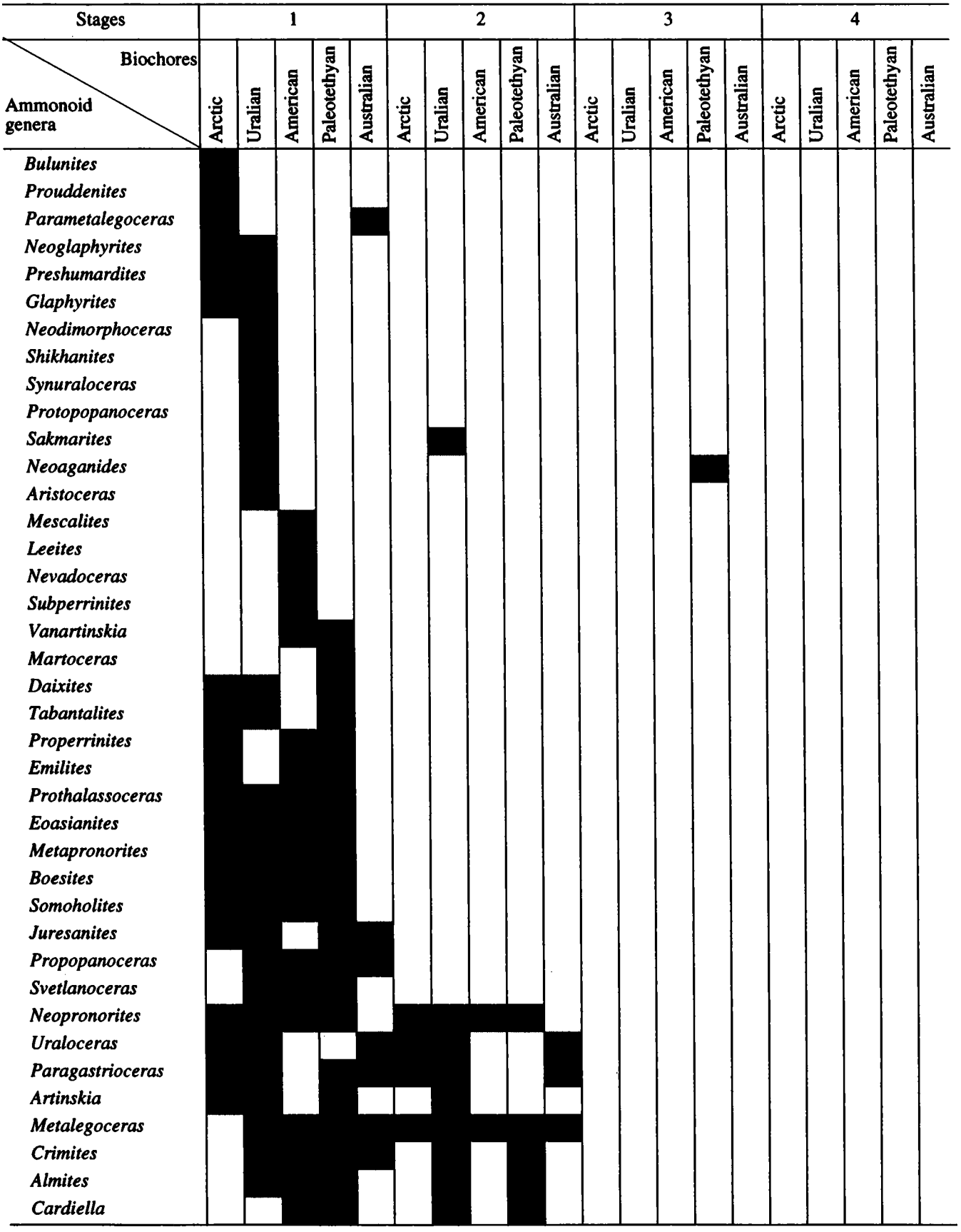
AMERICAN BIOCHORE

The American biochore extends throughout Central and North America, except for Alaska, Yukon, and Arctic Archipelago of Canada (Fig. 2). During the Early Permian, ammonoids of the American biochore were moderately diverse and represented 27% of all Asselian and 36% of all Sakmarian genera. Four of the 23 genera that existed in this period were endemic (*Mescalites, Subperrinites, Leeites, and Nevadoceras*). At the second stage, the ammonoid diversity decreased to 23% in the Artinskian Age and to 14% in the Kungurian time, whereas the relative abundance of endemic genera (*Stenolobulites, Akmilleria, Metacrimites, Perrinites, Pseudovidrioceras*, and *Peritrochia*) became greater. The Early Permian genus *Stacheoceras* is known only from American regions. Since the very beginning of the Late Permian, the taxonomic diversity of ammonoid genera was growing to 32 and 38% in the Roadian and Wordian ages, respectively; in the Capitanian time, they already represented over 50% of all known genera. According to the preferable viewpoint, the La Colorado Formation of Mexico is of the late Capitanian age and corresponds to the Amarassian deposits in Timor bearing genera *Episageceras, Timorites, Kingoceras, Stacheoceras, Xenodiscus*, and *Eoaraxoceras* (Cantu-Chapa, 1997). Ammonoids of the fourth stage are absolutely unknown in the American regions (Fig. 3c). It should be mentioned, in addition, that the Wordian and Capitanian ammonoid assemblages from America are similar to those from the Paleotethys, but their direct interrelations remain ambiguous. Apparently, some other factors might be responsible for this circumstance.

ARCTIC BIOCHORE

The Arctic biochore includes the ammonoid localities of northeastern Asia, the Novaya Zemlya Archipelago, the Pai-Khoi Range, Vaigach Island, the Canadian

Distribution of ammonoid genera in various biochores



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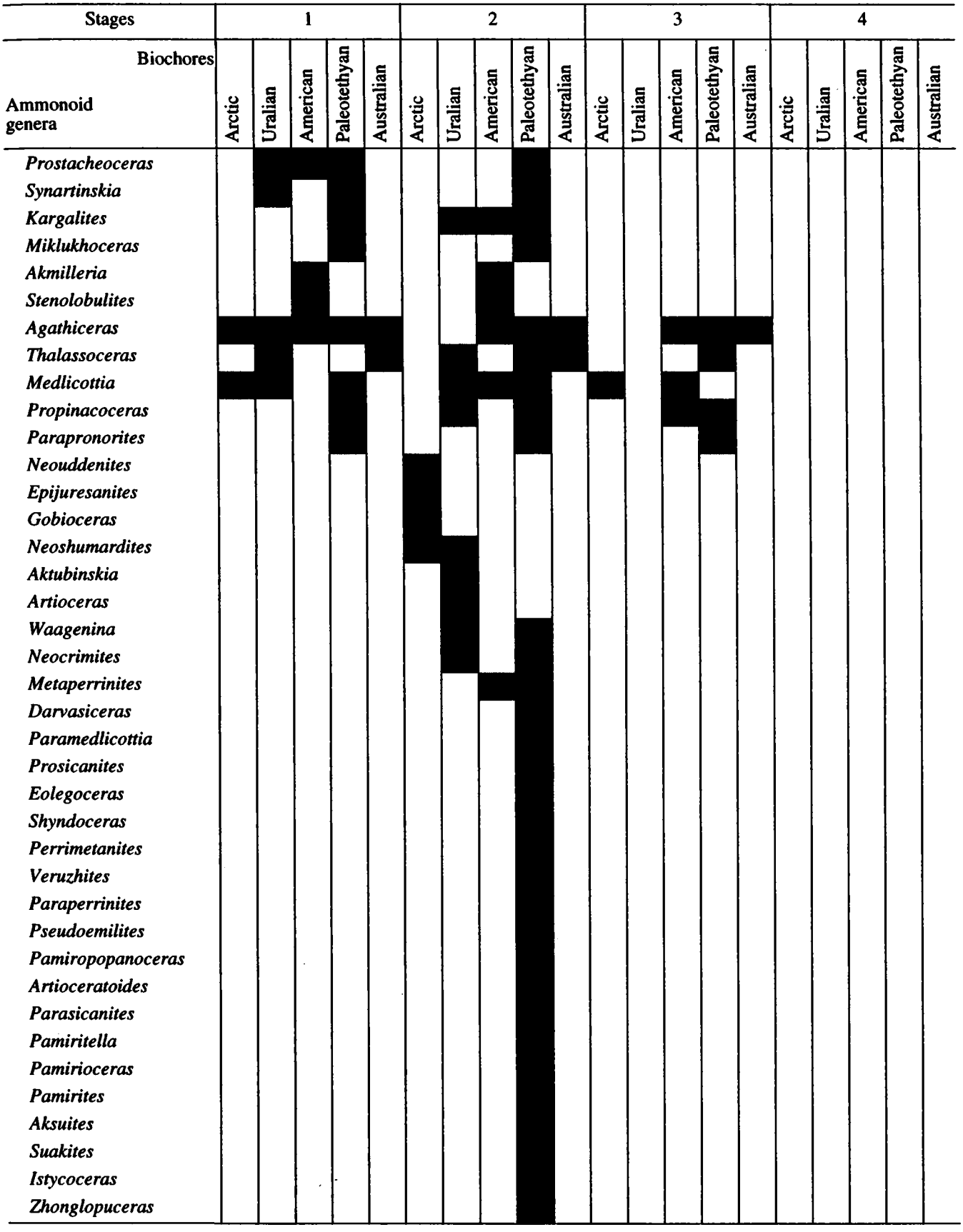
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Uralian
American
Paleotethyan
Australian
Prostacheoceras
Synartinskia
Kargalites
Miklukhoceras
Akmilleria
Stenolobulites
Agathiceras
Thalassoceras
Medlicottia
Propinacoceras
Parapronorites
Neouddenites
Epijuresanites
Gobioceras
Neoshumardites
Aktubinskia
Artioceras
Waagenina
Neocrimites
Metaperrinites
Darvasiceras
Paramedlicottia
Prosicanites
Eolegoceras
Shyndoceras
Perrimetanites
Veruzhites
Paraperrinites
Pseudoemilites
Pamiropopanoceras
Artioceratoides
Parasicanites
Pamiritella
Pamirioceras
Pamirites
Aksuites
Suakites
Istycoceras
Zhonglopuceras

Table. (Contd.)

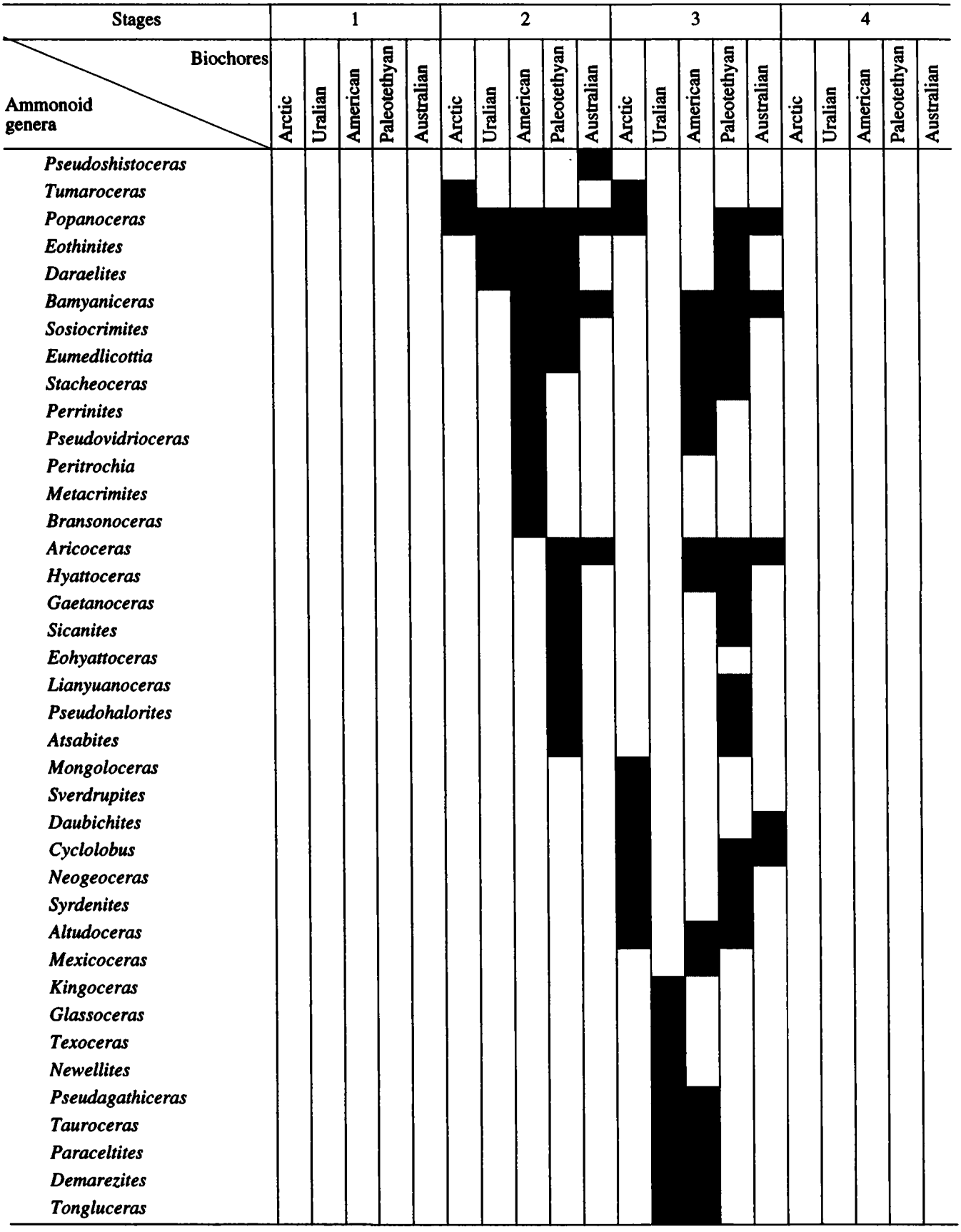
Stages
1
2
3
4
Biochores
Ammonoid genera
Arctic
Uralian
American
Paleotethyan
Australian
Pseudoshistoceras
Tumaroceras
Popanoceras
Eothinites
Daraelites
Bamyaniceras
Sosiocrimites
Eumedlicottia
Stacheoceras
Perrinites
Pseudovidrioceras
Peritrochia
Metacrimites
Bransonoceras
Aricoceras
Hyattoceras
Gaetanoceras
Sicanites
Eohyattoceras
Lianyuanoceras
Pseudohalorites
Atsabites
Mongoloceras
Sverdrupites
Daubichites
Cyclolobus
Neogeoceras
Syrdenites
Altudoceras
Mexicoceras
Kingoceras
Glassoceras
Texoceras
Newellites
Pseudagathiceras
Tauroceras
Paraceltites
Demarezites
Tongluceras

Table. (Contd.)

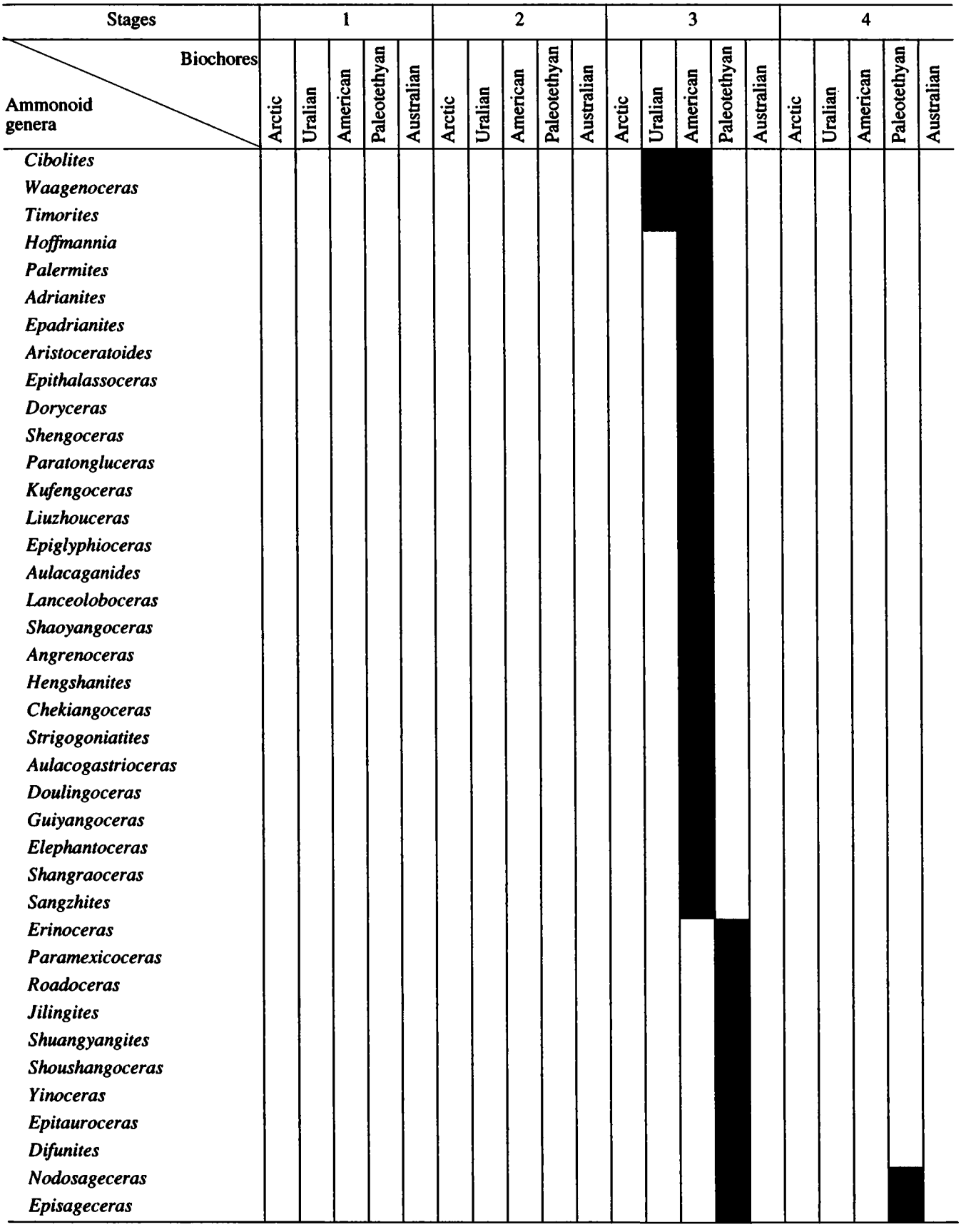
Stages
1
2
3
4
Biochores
Ammonoid genera
Arctic
Uralian
American
Paleotethyan
Australian
Cibolites
Waagenoceras
Timorites
Hoffmannia
Palermites
Adrianites
Epadrianites
Aristoceratoides
Epithalassoceras
Doryceras
Shengoceras
Paratongluceras
Kufengoceras
Liuzhouceras
Epiglyphioceras
Aulacaganides
Lanceoloboceras
Shaoyangoceras
Angrenoceras
Hengshanites
Chekiangoceras
Strigogoniatites
Aulacogastrioceras
Doulingoceras
Guiyangoceras
Elephantoceras
Shangraoceras
Sangzhites
Erinoceras
Paramexicoceras
Roadoceras
Jilingites
Shuangyangites
Shoushangoceras
Yinoceras
Epitauroceras
Difunites
Nodosageceras
Episageceras

Table. (Contd.)

Stages	1					2					3					4				
Biochores / Ammonoid genera	Arctic	Uralian	American	Paleotethyan	Australian	Arctic	Uralian	American	Paleotethyan	Australian	Arctic	Uralian	American	Paleotethyan	Australian	Arctic	Uralian	American	Paleotethyan	Australian
Eoaraxoceras																				
Xenodiscus																				
Araxoceras																				
Rotaraxoceras																				
Prototoceras																				
Pseudotoceras																				
Vescotoceras																				
Urartoceras																				
Dzhulfoceras																				
Vedioceras																				
Avushoceras																				
Periptychoceras																				
Anfuceras																				
Konglingites																				
Jinjiangoceras																				
Qinglongites																				
Kiangsiceras																				
Sanyangites																				
Pericarinoceras																				
Pachyrotoceras																				
Xiangulingites																				
Lenticoceltites																				
Huananoceras																				
Pseudogastrioceras																				
Phisonites																				
Iranites																				
Dzhulfites																				
Shevyrevites																				
Paratirolites																				
Abichites																				
Changhsingoceras																				
Liuchengoceras																				
Rongjiangoceras																				
Pentagonoceras																				
Rotodiscoceras																				
Qianjiangoceras																				
Pleuronodoceras																				
Pernodoceras																				
Sinoceltites																				
Pachydiscoceras																				
Dushanoceras																				
Chaotianoceras																				
Pseudotirolites																				
Mingyuexiaceras																				
Pseudostephanites																				
Trigonogastrites																				
Tapashanites																				

Arctic Archipelago, Yukon, Alaska, and Greenland (Fig. 2). Ammonoids of this biochore were very diverse in the initial Early Permian time. Their genera represented 45% in the Asselian and 44% in the Sakmarian ages relative to the total number of coeval taxa. In addition to endemic genera *Bulunites* and *Prouddenites*, there were widespread species of *Paragastrioceras, Uraloceras, Prothalassoceras, Eoasianites, Metapronorites, Boesites, Somoholites, Juresanites*, and some other genera. Spinosa *et al.* (1991) reported on an interesting "intermediate" ammonoid fauna from Nevada and Yukon, where it is distinct due to the coexistence of "Boreal" genera *Paragastrioceras* and *Uraloceras* with the "Tethyan" genus *Properrinites* and cosmopolitan forms. The map illustrating distribution of the *Paragastrioceras–Uraloceras* fauna in this work should be supplemented with data on the localities of Sakmarian ammonoids in the Pamirs (Darvaz) and Australia. However, it is necessary to mention that two *Paragastrioceras* forms known from the Darvaz locality (Leven *et al.*, 1992) are scarce and small, suggesting a depressed state of this fauna in the last region.

At the second stage, the ammonoid fauna of the biochore became less diverse, and abundance of genera fell to 16% in the Artinskian and to 11% in the Kungurian ages. In contrast, the extent of endemism increased, and four genera (*Neouddenites, Epijuresanites, Tumaroceras*, and *Gobioceras*) of all ten existed in this period were endemics (table, Fig. 4).

The third stage marks the new event of diversity decrease, and the abundance rate of Roadian, Wordian, and Capitanian taxa became 18, 2, and 12%, respectively (Fig. 3d). The Roadian ammonoid assemblage includes only 9 genera, one of which (*Sverdrupites*) was endemic. The ammonoid assemblage from Greenland comprises the genus *Cyclolobus*, characteristic as well of the Paleotethys, and the geographic position of Greenland in the Permian time remains disputable.

AUSTRALIAN BIOCHORE

The biochore in question (Fig. 2) includes the ammonoid occurrences known in the Sakmarian–Capitanian sediments of western Australia and in the Artinskian–Roadian deposits in the eastern part of the continent. Their diversity changed in the following succession. During the first stage (Sakmarian Age), there were 8 genera (with a relative abundance rate of 18%), and almost all of them (*Agathiceras, Thalassoceras, Paragastrioceras, Uraloceras, Juresanites, Metalegoceras*, and *Propopanoceras*) were cosmopolitan. Nine genera of the second stage represented from 16% (Artinskian Age) to 5.5% (Kungurian Age) of the total amount of taxa. The widespread genera coexisted with *Pseudoshistoceras* forms characteristic of Australia only. The third stage with 6 genera (Fig. 3e) shows the following abundance rates: 4% for the Roadian, none for the Wordian, and 3% for the Capitanian. The ammonoid community under consideration was very enigmatic: *Agathiceras* and *Popanoceras* forms were cosmopolitan; genera *Bamyaniceras* and *Aricoceras* represented taxa simultaneously known from the Paleotethys; and *Daubichites* was a characteristic member of Boreal communities. In addition, the Capitanian deposits of western Australia yield *Cyclolobus* forms known as well from eastern Greenland, the Salt Range, Tibet, and Madagascar.

CONCLUSION

(1) The evolution history of Permian ammonoids can be divided into four principal stages: "pro-Permian" (initial Early Permian), "eo-Permian" (terminal Early Permian), "meso-Permian" (initial Late Permian), and "meta-Permian." Each of the stages included ammonoid assemblages displaying their own trend of biogeographic differentiation.

(2) Analysis of regional ammonoid faunas revealed five principal biochores: Paleotethyan, Uralian, American, Arctic, and Australian, each populated by communities with different dominant and endemic genera.

(3) Changes of the taxonomic diversity in the ammonoid biochores are as follows:

a) at the beginning of the Early Permian, the proportion of genera in common was rather high in all biochores;

b) the end of the Early Permian is marked by a distinct differentiation of the ammonoid distribution areas, though the Paleotethyan regions of Pamirs, Afghanistan, southwestern China, Thailand, and Timor were interconnected during the entire Early Permian period, and the fauna interchange was most intense in the second half of the period;

c) the free interchange of pelagic faunas at the commencement of the Late Permian (especially during the Wordian time) was typical of the basins of Sicily, Tunisia, Oman, Iraq, Crimea, Greece, Croatia, and Timor; the coeval ammonoid faunas of America also included many Paleotethyan forms;

d) at the end of the Permian, the shallow basins of southern China, Far East, Japan, Iran, and Transcaucasia were populated by affiliated ammonoid communities, probably because of similar habitat environments in the closely spaced basins interconnected by migration ways.

ACKNOWLEDGMENTS

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Quaternary Loess–Paleosol Stratigraphy of Central Asia: Geochronology, Correlation, and Evolution of Paleoenvironments

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Abstract—Loess–paleosol covers are widespread in Central Asia. The most complete loess–paleosol sections are situated in southern Tajikistan. By using biostratigraphic and paleomagnetic records, we recognized the main stratigraphic units of the upper Pliocene and Pleistocene in the sections. The thermoluminescence analysis and correlation of loess–paleosol successions to the oxygen isotope scale are used to obtain the refined loess–paleosol geochronology of Central Asia. Thorough measurements of magnetic susceptibility (κ) of paleosol and loess horizons revealed structural peculiarities of individual soil beds. Data on climatic fluctuations inferred from the loess–paleosol successions are in accordance with the oxygen isotope records. The loess sedimentation is shown to be closely related to the atmospheric circulation and synoptic phenomena.

Key words: *loess, pedocomplex, stratigraphy, magnetic susceptibility, correlation, thermoluminescence*

INTRODUCTION

The loess–paleosol sequences represent the most complete geological record of succession, frequency and intensity of paleoclimatic fluctuations on land. The most significant information is inferable from buried soils and pedocomplexes (PC). Data on loess–paleosol successions enable us to elaborate the refined climatostratigraphic scales for continental deposits accumulated during the last one or two million years and to correlate them with the oxygen isotope records of paleoclimatic events in oceanic sediments (Kukla and An, 1989; Heller *et al.*, 1987; Shackleton *et al.*, 1995). This is a solid basis for considering problems of stratigraphy and the correlation of continental and marine deposits, and also the complex environmental evolution.

In the late 1970s, a stratigraphic scale for alluvial–proluvial and loess–paleosol successions was established by using data of biostratigraphic, paleopedological and thermoluminescence (TL) methods (Pen'kov *et al.*, 1976; Dodonov and Pen'kov, 1977). In parallel, the loess–paleosol stratigraphic scale was suggested (Lazarenko *et al.*, 1977). The alluvial–proluvial sequences and lithostratigraphic complexes bearing the late Pliocene–early Pleistocene mammalian fauna were correlated with the old loess–paleosol formations. The loess-forming processes began 2.5 Ma ago, and the Matuyama–Brunhes paleomagnetic reversal was shown to be recorded between pedocomplexes 10 and 9. Within the Brunhes Chron, ages of loess and paleosol horizons were determined on the basis of TL dating by V.N. Shelkoplyas from the Kiev Thermoluminescence Laboratory. Because of a wide range of the TL dates obtained and in accordance with different stratigraphic interpretations of some paleosol horizons, two alternative loess–paleosol stratigraphic schemes for the Brunhes interval were worked out. The last interglacial was considered as corresponding either to the third pedocomplex (PC3) (Lazarenko *et al.*, 1977) or to the fourth pedocomplex (PC4). The scheme used here for numbering buried soils and pedocomplexes is that suggested by Lazarenko (1984) and Bronger *et al.* (1993). It should be mentioned that the results obtained at the Kiev Thermoluminescence Laboratory date the Paleolithic sites of PC5 (Karatau) and PC5–PC4 (Lakhuti and Obimazar) back to 200 and 130 ka, respectively (Lazarenko and Ranov, 1977; Dodonov *et al.*, 1978). New data presented here allow us to revise these dates.

The development of the thermoluminescence technique during the last 15 years and close check of the TL dating by other methods revealed that the precision and, hence, the reliability of the TL dating are valid within the time range of 100–150 ka (Wintle, 1987). This made us be severe on previous TL dates for loess beds in Central Asia where, as we suppose, those for the PC4 and PC5 bearing archaeological artifacts should be revised. In addition, the study on the Loess Plateau in

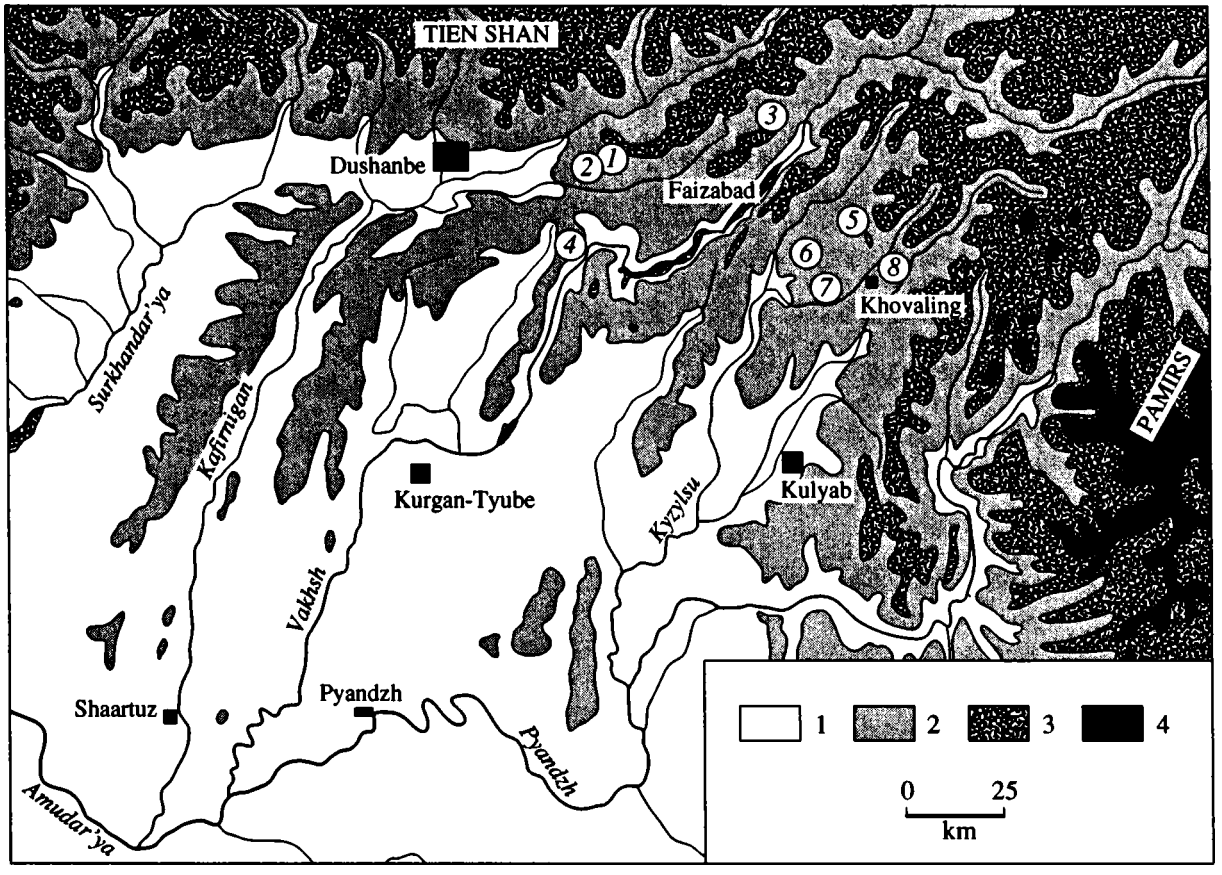

Fig. 1. Hypsometric map of southern Tajikistan and locations of the Upper Pliocene–Quaternary reference sections. Hypsometric scale: (1) 1000 m, (2) 1000–2000 m, (3) 2000–4000, (4) >4000 m. Geological sections (encircled figures): (*1*) Karamaidan, (*2*) Zil'fi, (*3*) Obigarm, (*4*) Karatau, (*5*) Darai Kalon and Chashmanigar, (*6*) Kuruksai, (*7*) Lakhuti, (*8*) Khonako.

northern China, where the loess–paleosol sequences are very similar in structure to those of Central Asia, showed that the first paleosol complex (S1) of China ranges in age from 130 to 70 ka and can be correlated to the oxygen isotope stage 5 (Liu *et al.*, 1985; Kukla *et al.*, 1990). However, the first well-developed pedocomplex (PC1) in southern Tajikistan has been previously identified as being much younger in age. This was an obstacle for the correlation of the loess–paleosol sequences in Central Asia and China that stimulated comprehensive stratigraphic investigations of the respective rock sequences in Central Asia, especially in southern Tajikistan (Fig. 1), where the subaerial sedimentary covers are very thick and complete.

RECENT CONDITIONS AND CLIMATIC FACTORS OF DUST ACCUMULATION

The arid zones of the world produce enormous dust masses experiencing transportation, sedimentation, and initial lithogenetic transformation. In Central Asia, the summer–autumn period is favorable for dust storms owing to the minimal precipitation and intense drying of soil covered by scarce vegetation. In general, these factors promote deflation. The number of dust-storm days per year ranges from five in the Pamir–Alai and Tien Shan valleys to 50 and more in Turkmenistan and other areas of sandy grounds uncovered by vegetation (Fig. 2). Synoptic conditions typical of the dust storm development during the warm period in Central Asia are related to the penetration of cold atmospheric masses from the west, northwest, and north. When speaking about the influence of cold air masses, we should mention a synoptic phenomenon characteristic of the southern parts of Uzbekistan and Tajikistan, such as the strong dry wind from the west and southwest, the so-called "Afghan" responsible for deflation in the desert and semidesert areas near the northern border of Afghanistan. Anywhere from 10 to 30 dust storms per year occur in these areas. Researchers, who carried out observations in southern Tajikistan according to the joint Soviet–American project of experimental studies of the arid aerosol, recorded the wind velocity approaching 28 m/s in Aivadzh (with gusts up to 34 m/s) and 18–20 m/s in Shaartuz during the dust storm on September 20, 1989 (Krapivtseva *et al.*, 1992). The dust mist risen by frontal wind spread over the Kurgan-Tyube, Kulyab, and other areas of southern Tajikistan.

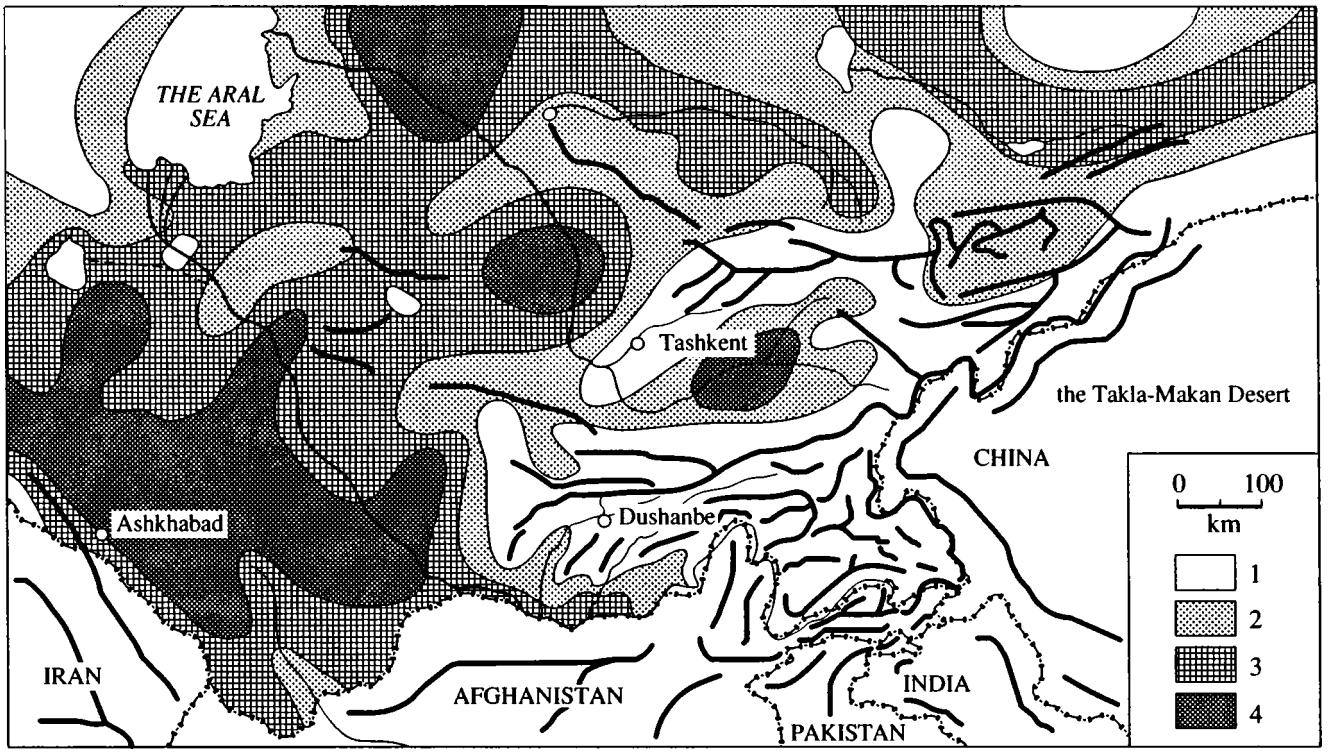

Fig. 2. Annual duration of dust storms in Central Asia: (1) less than 5 days, (2) 5–10 days, (3) 10–30 days, (4) 30–50 days and longer.

The field of vision was reduced to 0.2 km in Aivadzh and to 0.5 km in Shaartuz, Esanbai, and Dushanbe. The dust mist rose to more than 3 km, as recorded in the Sanglok, Anzob and Bustonobad observatories, situated at 2200–3300 m above sea level. The altitude limit of the dust aerosol dispersion is largely controlled by the temperature inversion producing a "blocking" effect that notably decreases the coefficient of dust aerosol dispersion over the inversion layer. The "Afghan" transfers enormous dust masses from the south to the north and northeast along the Surkhandar'ya, Kafirnigan, Vakhsh, and Kyzylsu river valleys. High (4000–5000 m) ridges of the Tien Shan and West Pamirs serve as orographic barriers causing the stagnation of dust-saturated air flows and dust accumulation.

Another phenomenon characteristic of Central Asia is the dusty atmosphere (dust mist). When there are no dust storms, the dust rises because of intense convective air flows in the daytime during the summer thermal depressions. Nightly, dust is accumulated much slower than it rises during the daytime convection. This increases the mist density during the hot period. In southwestern Tajikistan, the longest period of the dust mist is the July–October season (*Spravochnik...*, 1970). The end of October is when the rainy season begins. During the winter, owing to the higher precipitation and moistening of the desert surface, the dust accumulation is sharply reduced and the mist period become shorter.

Theoretically estimated masses of dust aerosol per unit of the air volume, the altitude range of dust dispersion, and the rate of dust accumulation show that the annual dust accumulation can vary from 0.04 to 0.683 mm between different areas of southern Tajikistan; the mean value is 0.2 mm for the entire region (Finaev, 1995). According to experimental measurements during the dust storms in Shaartuz on September 16 and 20, 1989, the dust accumulation was 41.1 g/m^2. Taking into account the average annual number (10–30) of dust storms in the region and the mean volume of accumulated dust, the rate of dust accumulation can be as high as 206–617 g/m^2 per year (Gillet and Dobrowolski, 1992), or 0.08–0.23 mm/year in terms of the loess specific weight (2.6 g/cm^3). However, these values do not show the actual intensity of sediment accumulation in subaerial environments, because the recurrent erosion and deflation affect the ratio between volumes of accumulated dust and preserved sediment. The relatively high rate of dust accumulation is incompatible with the relatively small loess thickness in the southern Afghan–Tajik depression, because the dust material is affected by eolian processes in this deflation area.

THE RECENT SOIL COVER

The recent soil cover in Central Asia is formed under different moisture, vegetation and relief condi-

tions thus being dependent on altitudinal zoning. Southern Tajikistan, where most of the reference loess–paleosol sections are located, demonstrates the main trends in the recent soil development above the loess substratum.

The lower relief zone between 300 and 700 (1000) m above the sea level is characterized by widespread sierozem. This zone is distinguished by the steppe vegetation and annual precipitation of 300–500 mm. In the sierozem profile, the carbonate content changes from 6–15 to 20–22% from the top downward (Sadriddinov, 1982). At a depth of 30–60 cm, the sierozem profile usually contains abundant bioliths consisting of separate capsules and chambers produced by invertebrate pedofauna, whose larvae build nests with thickened walls to pupate under the dry summer conditions (Valiakhmedov, 1977). Buried capsules and chambers are filled with enclosing deposits.

In the piedmont zone (700–2500 m), there are widespread brown soils. The lower steppe part of this zone is covered with xerophytic shrub vegetation replaced higher by broad-leaved forests considerably reduced because of the human activity. The following subtypes of brown soils can be recognized: the carbonate brown soil in the altitude interval of 700–1000 (1400) m, where the annual precipitation is 500–700 mm/year, typical brown soil in the interval of 1000 (1400)–1800 m with the annual precipitation of 700–900 mm/year, and leached brown soil in the interval of 1500–2500 m with the annual precipitation of 1000–1200 mm/year. Profiles of different subtypes show different carbonate distribution patterns. For instance, brown carbonate soils are rich in carbonates within the upper part of their profile, while the illuvial–carbonate horizon is situated at the depth of 40–50 cm in the typical brown soils and deeper than 50–60 cm in the leached brown soils.

At an altitude of 1500–2000 (2200) m, a fragmental distribution of forest brown soils is recorded within limited areas of broad-leaved forests, where the annual precipitation exceeds 1000 mm. Clayey illuviation occurs here at the depth of 40–50 cm and deeper, and the Bt horizon may be as thick as 60–80 cm. The basal horizon Cca is characterized in this case by an abundance of carbonate concretions (Lomov, 1985).

LOESS–PALEOSOL STRATIGRAPHY

Loess deposits are widespread in the piedmonts of Tien Shan and the Pamirs. They form an almost continuous cover above river terraces, watersheds, and ridge slopes. The loess distribution is hypsometrically limited by the absolute height of 2000–2500 m, although fragments of thin loess cover occur even higher (up to 3500–4000 m), e.g., in the Alai valley and East Pamir Plateau. In the watershed areas of upper reaches of the Kafirnigan, Vakhsh and Kyzylsu rivers of southern Tajikistan, the maximal thickness of the loess–paleosol sequences is 100–200 m. The best studied reference loess–paleosol sections are located at the Karamaidan, Chashmanigar, Darai Kalon, Lakhuti, and Khonako sites there (Fig. 1). In the river valleys and depressions, the loess–paleosol facies are replaced by alluvial–proluvial deposits represented by gravels, sands, silts, and loams. The mammalian remains from alluvial–proluvial deposits are divided into two heterochronous faunal assemblages attributed to the middle Villafranchian (the Kuruksai, Obigarm, Karamaidan, and Zil'fi localities) and the Cromerian (the Lakhuti-2 locality) (Vangengeim *et al.*, 1988; Sotnikova, 1989; Dodonov *et al.*, 1991). According to paleontological and paleomagnetic data, three main stratigraphic units can be distinguished in the subaqueous and subaerial deposits: the upper Pliocene (the Kuruksai Formation), Eopleistocene (the Kairubak Formation), and Pleistocene (the Kyzylsu Group).[1] The oldest loess beds and buried soils show the reversed magnetic polarity corresponding to the Matuyama Chron. They are of the late Pliocene–Eopleistocene age in the Karamaidan, Zil'fi, Obigarm, Chashmanigar, and Khonako sections. In all loess–paleosol sections, the Matuyama–Brunhes reversal was established between the PC10 and PC9. The Jaramillo Subchron corresponds to the PC15 and PC16 of the Chashmanigar and Khonako sections, and the Olduvai Subchron, which was reliably recognized only in the Chashmanigar section, spans the PC33, PC34 and the most part of PC35 (Dodonov and Pen'kov, 1977; *Putevoditel'...*, 1977). The Karamaidan section shows the largest number (40–43) of buried soil horizons (pedocomplexes) and the Gauss–Matuyama reversal located in its lowermost part. However, it should be noted that the Karamaidan section is composite and unites at least three conventionally correlated sections. Therefore, the exact number of buried soils needs to be defined.

In the last 10–15 years, the lower parts of some sections became inaccessible because of landslides. For instance, large landslides covered the oldest beds of the Chashmanigar and Khonako sections. One of the most complete loess–paleosol sections, being 176 m thick, is that of the Darai Kalon site located in the neighboring landslide area, 400 m to the south of the Chashmanigar Exposure (Fig. 3). This section exhibits 18 horizons of paleosols and several pedocomplexes. The pedocomplexes are composed of two or three paleosol horizons illustrating different types of soil-forming process. The Darai Kalon loess–paleosol section, when supplemented with data on other sections, has the following characteristics. The first pedocomplex (PC1) consists of three closely spaced soils; the lowest typical brown soil, the middle one attributed to the carbonate brown soil, and the upper one of the sierozem type. The PC2 has the two-member structure, where the lower member ($PC2_2$) is a typical brown soil and the upper one

[1] The Eopleistocene corresponds to the lower Pleistocene, and the Pleistocene to the middle–late Pleistocene of the West European scale.

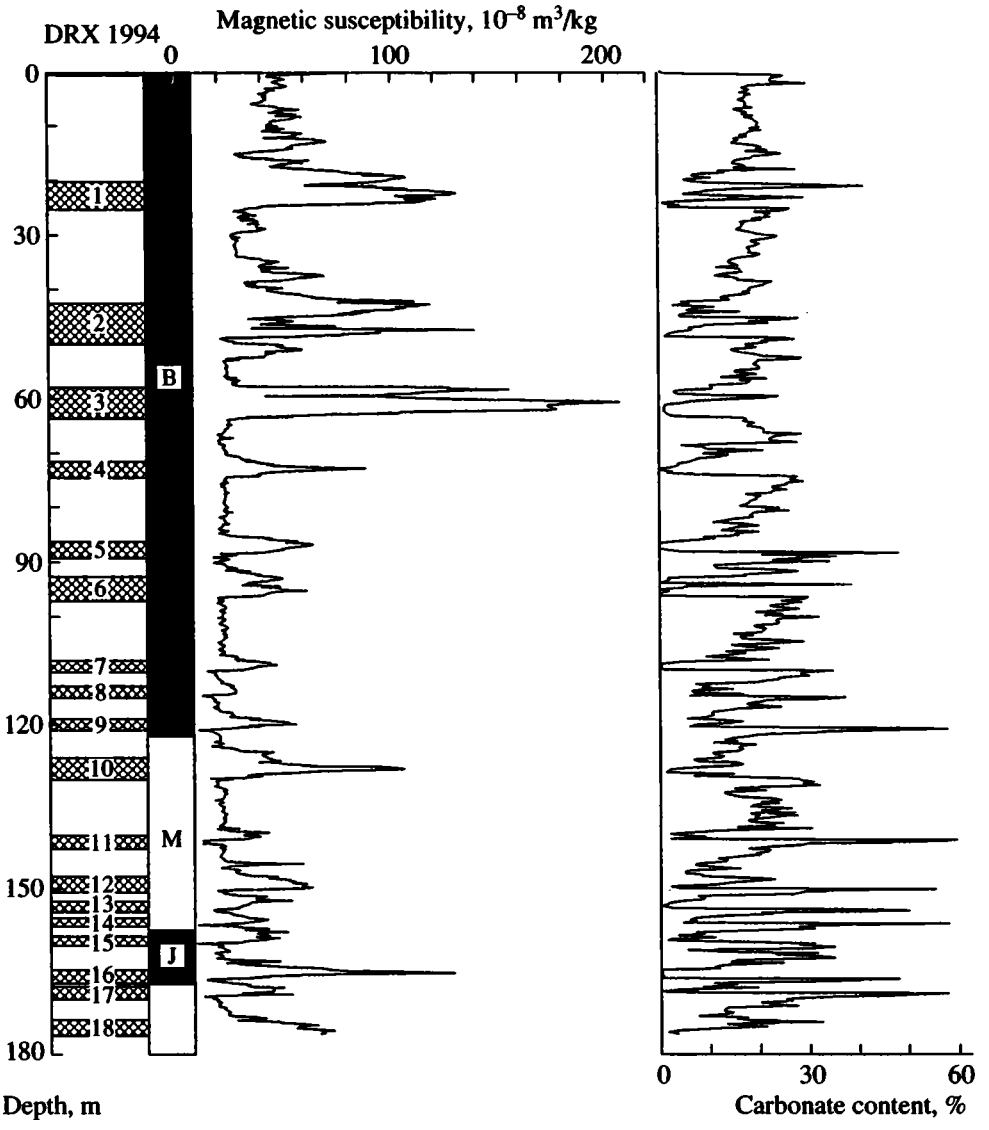

Fig. 3. Magnetic susceptibility and calcium carbonate content across the Darai Kalon loess–paleosol section. Magnetic polarity of rocks is shown in accordance with the magnetic scale of the adjacent Chashmanigar section (black and white intervals denote the normal and reversed polarity, respectively): Brunhes Chron (B), Matuyama Chron (M); (J) Jaramillo Subchron; (1–18) pedocomplexes.

($PC2_1$) is similar to the carbonate brown soil. The analogous structures of PC1 and PC2 are recognized in all the sections studied. At the Lakhuti site, these pedocomplexes overlap each other. In the eastern part of the Karamaidan Exposure, the PC1 is considerably eroded. PC3 consists of two well-developed and weakly leached brown soils. The lower bright red–brown soil has an extended profile and a high clay content. Morphological analysis of three upper pedocomplexes (PC1–PC3) revealed their complex structure reflecting several stages of soil formation. It is characteristic that the lowermost soil of each pedocomplex was formed under relatively wet climatic conditions responsible for origin of the typical brown soils.

The PC4 is characterized by a compact profile displaying signs of clayey illuviation at the Bt horizon. The latter is represented by a reddish-brown loam showing the prismatic structure and ferrous–manganese films along the joints. In its micromorphological aspect, the Bt horizon is characterized by clayey cutans and decalcification. Thus, the most developed part of the PC4 profile can be classed with the leached or mountainous–forest brown soil. The lower part of PC4 includes the distinct illuvial–carbonate Cca horizon 30–40 cm thick.

In the general structure and features characterizing the most developed part of the profile, the PC5 is very similar to PC4. Its main member attributed to the leached brown soil shows distinct signs of secondary

carbonatization. Micromorphological signs of clayey illuviation are less distinct in PC5 than in PC4. The illuvial–carbonate horizon of PC5 is slightly thicker (50–80 cm) than that of PC4. The PC6 consists of two soils. The illuvial–carbonate horizon of the upper soil overlaps the lower one. The upper soil ($PC6_1$) is of the carbonate brown type, whereas the lower soil ($PC6_2$) includes the Bt horizon, whose lower part is leached, and we consider this soil to be of the leached brown type. The PC7 is relatively simple in structure. In the Darai Kalon and Khonako sections, it includes the Bt horizon and can be referred to the leached brown soils, whereas the same unit represents the carbonate brown soil in the Karamaidan section. The PC8 and PC9 have the simple structure of the carbonate brown soils. The PC10 consists of two paleosols: the lower one ($PC10_2$) corresponding to the weakly leached brown soil and the upper soil ($PC10_1$) of the carbonate brown type. The both paleosols show distinct illuvial–carbonate horizons.

According to analytical data, the carbonate content in the loess horizons varies within from 15–25% (Fig. 3). In the Darai Kalon section, the upper parts of horizons L2, L3, L4, L5, and L7 are quite rich in carbonate components. In general, the higher carbonate content in loess suggests highly arid climatic conditions and the desert–steppe environment. Occurrences of scattered concretions and their concentrations in the loess horizons indicate fluctuations in the moisture regime during the loess accumulation. The PC1–PC3 soils are characterized by different carbonate content, and their most leached varieties are situated in the lower parts of the pedocomplexes. The well leached soils (0–1% of $CaCO_3$) of PC4, PC5, PC6, PC7, and PC10 evidence for the relatively humid conditions of their formation. The PC8 and PC9 show an increase (to 5 or 6%) of the carbonate content. $CaCO_3$ content is as high as 50–60% in the illuvial–carbonate horizons of paleosols, which is particularly characteristic of soils in the lower part of the section (Fig. 3).

The general thickness of the loess–paleosol sequence, incorporating ten pedocomplexes, varies in different parts of the Afghan–Tajik depression because of different rates of loess accumulation and erosion. For instance, it is 67 m in Karamaidan, about 100 m in Khonako, and 130 m in the Darai Kalon sections. Significant thickness variations of individual loess horizons are evidently a result of erosion. For instance, the uppermost loess horizon L1 is anomalously thin (3–4 m) in the Karamaidan section, being 19 and 21 m thick in the Darai Kalon and Lakhuti sections, respectively. It is significant that the middle part of the loess horizon L1 in the Darai Kalon section shows signs of soil-type reworking. On the basis of observable pedofaunal tracts and frequent bioliths, this 7 to 8 m thick horizon can be referred to the sierozem. The extended thickness of the sierozem profile is unambiguous evidence of its synsedimentary formation. In general, the loess horizons alternating with ten upper pedocomplexes vary in thickness from 7 or 8 to 10–12 m. The exceptions are two loess horizons between the PC5 and PC6 and between the PC8 and PC9, where their thickness does not exceed 3–5 m.

Eight additional pedocomplexes (PC11–PC18) were recognized in the Darai Kalon section below the PC10. The B horizons of PC11, PC13, PC14, and PC17 are composed of massive brown, slightly calcareous loams lacking any signs of clayey illuviation. According to these features, these pedocomplexes can be defined as the carbonate brown soils. The PC12 and PC15 are similar to the listed units, but they include the relatively leached B horizons. High clay content, decalcification, a well-developed prismatic structure and red color are characteristic features of the B horizons in PC16 and PC 18. The red color may reflect the presence of hematite formed due to the relatively high seasonal temperature. The PC16 and PC18 can be conventionally attributed to the red earth of subtropical areas.

The loess horizons below PC11 are not more than 3–5 m thick. Some of these horizons, e.g., L13, L15, and L16, show rather high carbonate content.

MAGNETIC SUSCEPTIBILITY OF LOESS–PALEOSOL SEQUENCES

First measurements of magnetic susceptibility (κ) in the loess–paleosol sequences of southern Tajikistan revealed a distinct difference between the buried soils and loess beds: the magnetic susceptibility is high in the soils and low in the loess horizons (Dodonov and Pen'kov, 1977; Lomov and Pen'kov, 1979). The magnetic susceptibility of buried soils was found to be complicatedly dependent on the age and conditions of their formation. Further investigations in accordance with the INTAS project yielded more precise values of magnetic susceptibility of recent and buried soils. Sampling intervals were from 10 to 15 cm. Loose powdery samples dried at 40–50°C were used for magnetic susceptibility measurements in the stationary sensor MS2 "Bartington".

The magnetic susceptibility of recent soils varies widely, from 30 to 185×10^{-8} m^3/kg, whereas sierozems yield low κ values ($30–60 \times 10^{-8}$ m^3/kg). In profiles of the latter, the maximal κ values are typical of the humic horizon. Higher κ values ($50–185 \times 10^{-8}$ m^3/kg) are recorded in the brown soils, where they increase in the A horizon. Recent mountainous–forest and other brown soils have a similar magnetic susceptibility. Magnetic investigations of different soil types revealed the dependence of magnetic susceptibility on many factors, such as the soil type, composition of newly formed ferrous minerals, humus content, biogenic activity, introduction of strong ferromagnetics from the outside, transformation of ferrous minerals as a result of parent rock weathering, and, finally, the fire action (Babanin *et al.*, 1995). Studies on magnetic properties of soils in different landscape zones of Tajikistan showed that the κ value is

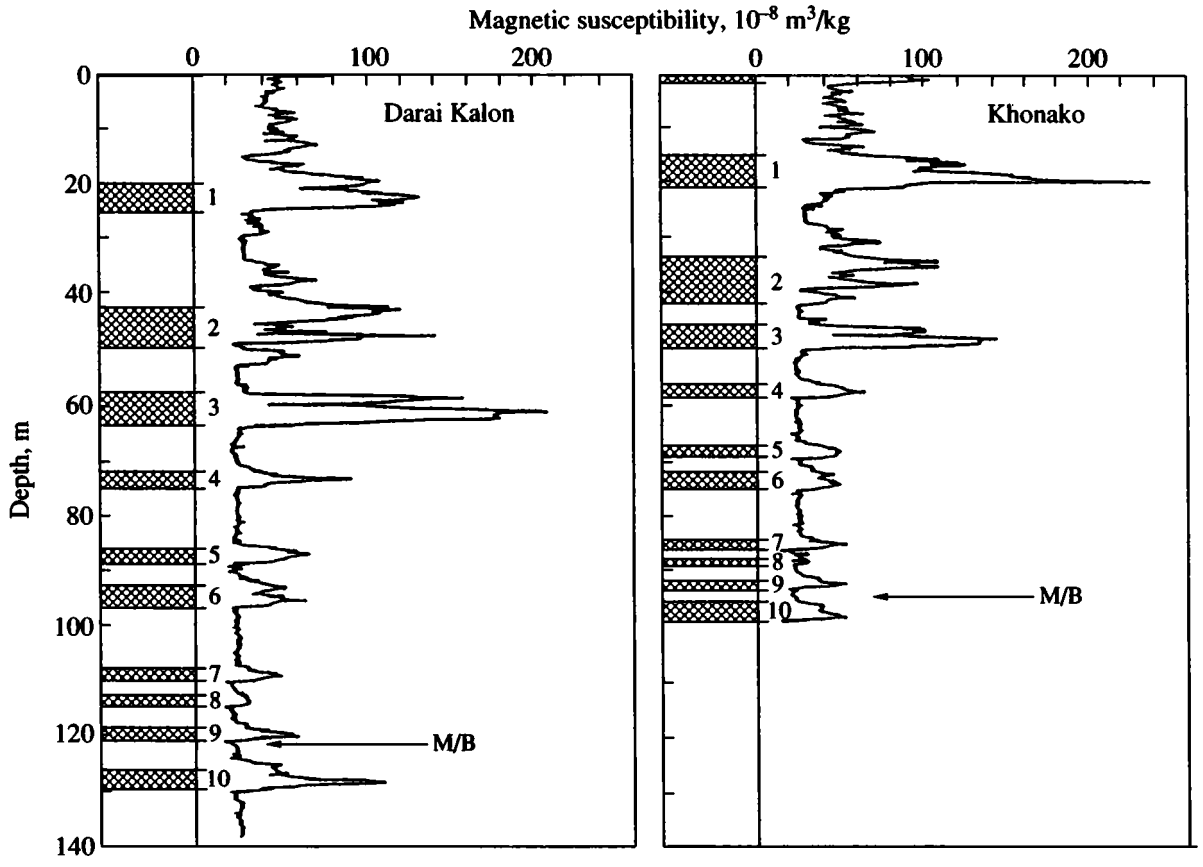

Fig. 4. Magnetic susceptibility (κ) of rocks in the Darai Kalon and Khonako loess–paleosol sections: (M/B) Brunhes–Matuyama reversal; (1–10) pedocomplexes.

controlled not only by the pedogenetic factors, but also by the mean annual temperature and relative moisture content. For instance, the high κ values are measured in piedmont soils of the Tien Shan and Pamirs, where the mean annual temperature is 8–10°C and precipitation is 800–1000 mm. It should be noted that the κ measurements may not be meaningful in the recent strongly eroded soils. For example, the recent soils of the Karamaidan, Lakhuti, and Darai Kalon sections were formed under climatic conditions favorable for the magnetic susceptibility, but being noticeably eroded, they yielded lower values. The analytical data obtained on recent soils can be used to model the paleoclimatic fluctuations on the basis of the loess–paleosol succession and the magnetic susceptibility of buried soils and loess horizons.

Values of the magnetic susceptibility in the loess horizons vary minimally, from 20–25 × 10^{-8} m^3/kg, though they increase to 30–40 × 10^{-8} m^3/kg in the immature sierozem in the middle part of the horizon L1 (Figs. 3, 4) and in the horizon L2, few meters above the PC2. In contrast to loess beds, the buried soils show high κ values. The κ peaks (up to 140 × 10^{-8} m^3/kg and higher) characterize the soil horizons of PC1, PC2, and PC3; the maximal value (235 × 10^{-8} m^3/kg) is obtained for the paleosol $PC3_2$ of the Darai Kalon section. It should be mentioned that the PC3 has been previously considered as the regional marker on the basis of high κ values measured in this unit by Pen'kov (Dodonov and Pen'kov, 1977; Lomov and Pen'kov, 1979).

The κ parameter decreases by transition from the upper pedocomplexes PC1–PC3 to the underlying PC4–PC18 (Figs. 3, 4). The latter profiles exhibit distinct illuvial–carbonate horizons probably indicating the seasonal moistening, which resulted in the formation of the Cca carbonate crusts. In periods of moistening, strong ferromagnetics, first of all, finely dispersed magnetite could be destroyed or their accumulation was impossible. This may be a factor lowering the κ parameter. In contrast, soils of the PC1–PC3 interval are more automorphous. It should be noted that the automorphous soils formed on loose deposits usually have higher κ values (Babanin *et al.*, 1995).

GEOCHRONOLOGY AND CORRELATION

Problems of geochronology and correlation of the loess–paleosol succession in Central Asia stem from inadequate interregional correlations based on TL dates previously obtained by Shelkoplyas at the Kiev Thermoluminescence Laboratory (*Putevoditel'...*, 1977). Here we discuss ages of the upper loess–paleosol hori-

zons within the Brunhes Chron, whereas older loess–paleosol sequences are well dated on the basis of paleomagnetic and biostratigraphic records. At the initial stage of investigations, the TL dates obtained at the Kiev Thermoluminescence Laboratory were important for the general idea of time succession of the loess horizons and buried soils. However, the TL dating method is shown to be of a limited usage, because the TL ages obtained for deposits older than 100–150 ka have, as a rule, been underestimated (Wintle, 1987, 1990; Forman, 1989). This is why the loess–paleosol geochronology of some European regions was revised in the last decade. For instance, the loess–paleosol geochronology of Hungary is revised on the basis of repeated TL dating (Bronger *et al.*, 1987; Singhvi *et al.*, 1989). A series of new TL measurements is used to date more precisely the buried soils and loess horizons in the middle part of the Rhine valley (Frechen, 1992). Additional TL dating of loess sequences is also carried out in northern France and Poland.

In accordance with objectives of the INTAS project, the experimental thermoluminescence measurements were made in the Orkutsai–Charvak section of the Tashkent region, rocks of which were previously investigated in several thermoluminescence laboratories (Punning *et al.*, 1992). The experimental measurements carried out in the Godwin Laboratory for Quaternary Research of Cambridge University showed that the TL ages older than 100–130 ka for the deposits below PC1 have been underestimated. It was also established that the age of the middle part of the section does not increase, and the TL date of 108.7 ± 13.6 ka at the level of Matuyama–Brunhes reversal is obviously underestimated (Zhou *et al.*, 1995). So, the opinion to consider as underestimated the TL ages for PC2, PC3, PC4, and PC5 appears to be justified. It should be mentioned that the previous TL dates (*Putevoditel'...*, 1977) for PC3 and PC4 (we do not recommend to use them now) were not older than 130–150 ka and the age of PC5 was estimated to be about 200 ka.

The TL measurements in a representative rock series from the Darai Kalon section at the Thermoluminescence Laboratory of Cologne University revealed the regular increase of TL dates in the upper loess horizon (L1). As established, the top of the immature sierozem paleosol within the horizon L1 is between dates 34.0 ± 0.2 and 26.3 ± 1.8 ka. The top of PC1 is found to be 57.3 ± 3.3 ka old, and the date of 96.7 ± 9.1 ka is measured in the middle part of PC1, whereas the loess horizon below PC1 is marked by dates >117 ka (Frechen and Dodonov, 1998). The TL dates become insignificantly older in the middle part of the Darai Kalon section and are not older than 300 ka at the level of the Matuyama–Brunhes reversal. These results support the inference from the experimental TL measurements in the Orkutsai section that the ages of loess horizons below the PC1 were earlier significantly underestimated. We do not consider here the factors responsible for this effect, because they are discussed in a number of publications on the technique of the TL measurements (Wintle, 1990; Forman, 1989; Singhvi and Mejdahl, 1985).

The TL measurements in rock samples from the Darai Kalon section at the Godwin Laboratory of Cambridge University yielded the dates of 68.9 ± 9.3 ka for the PC1 top and 119 ± 15.0 ka for the loess top below PC1 (Zhou, personal communication, 1995). The ages are well consistent with the TL dates obtained at Cologne Laboratory.

The magnetic susceptibility variation curves plotted in detail appear to be most helpful in the regional correlation of loess–paleosol sequences, especially when the soil horizons are overlapping or missing from a discontinuous section because of hiatus in sedimentation, and, finally, when we must correlate some fragmentary sections. The κ curves for the Darai Kalon, Khonako and Karamaidan sections characterize ten pedocomplexes within the Brunhes Chron and provide grounds for their unambiguous correlation.

The magnetic susceptibility curves for the loess–paleosol sequences are well correlative with the SPECMAP curve compatible with the oxygen isotope records when they are corrected on the basis of astronomical data on solar radiation (Imbrie *et al.*, 1984). The correlation of κ curves for loess sequences in Central Asia and China with the SPECMAP curve shows that the sequence of buried soils corresponds well to warm $\delta^{18}O$ stages (Fig. 5). Using as age markers the PC1 horizon corresponding to the last interglacial and the position of Matuyama–Brunhes reversal at 780 ka, we suggest the age calibration of the loess–paleosol scale of Central Asia for the last 800 ka (Table 1) on the basis of chronological succession of oxygen isotope stages coordinated with the SPECMAP curve and data from the ODP Site 677 (Bassinot *et al.*, 1994). The underdeveloped paleosol PC8 is correlative to the oxygen isotopic substage 18.3 $\delta^{18}O$ and to the soil horizon S7a of China. In loess–paleosol sections of Central Asia, the Matuyama–Brunhes reversal is one or two meters below PC9, whereas in the oxygen isotope scale it corresponds to the lower part of the warm stage 19. In Chinese sections, this reversal is within the loess horizon L8 (Heller *et al.*, 1987). If the loess sedimentation rate is 0.29 mm/yr within the horizon L10, the loess bed one or two meters thick may be accumulated during 3.5–7 thousand years. The estimated values illustrate a possible difference in age limits of climatic events near the Matuyama–Brunhes reversal, when these limits are calculated in accordance with the oxygen isotope and loess–paleosol scales, provided that the reversal has a stable position. However, epigenetic processes and secondary magnetization may displace the reversal recorded in subaerial deposits and deep-sea sediments to a somewhat lower position (Bol'shakov and Svitoch, 1996; Forster and Heller, 1994; Zhou and Shackleton, in press). On the other hand, the age limits of climatic events may slightly vary in different regions of the

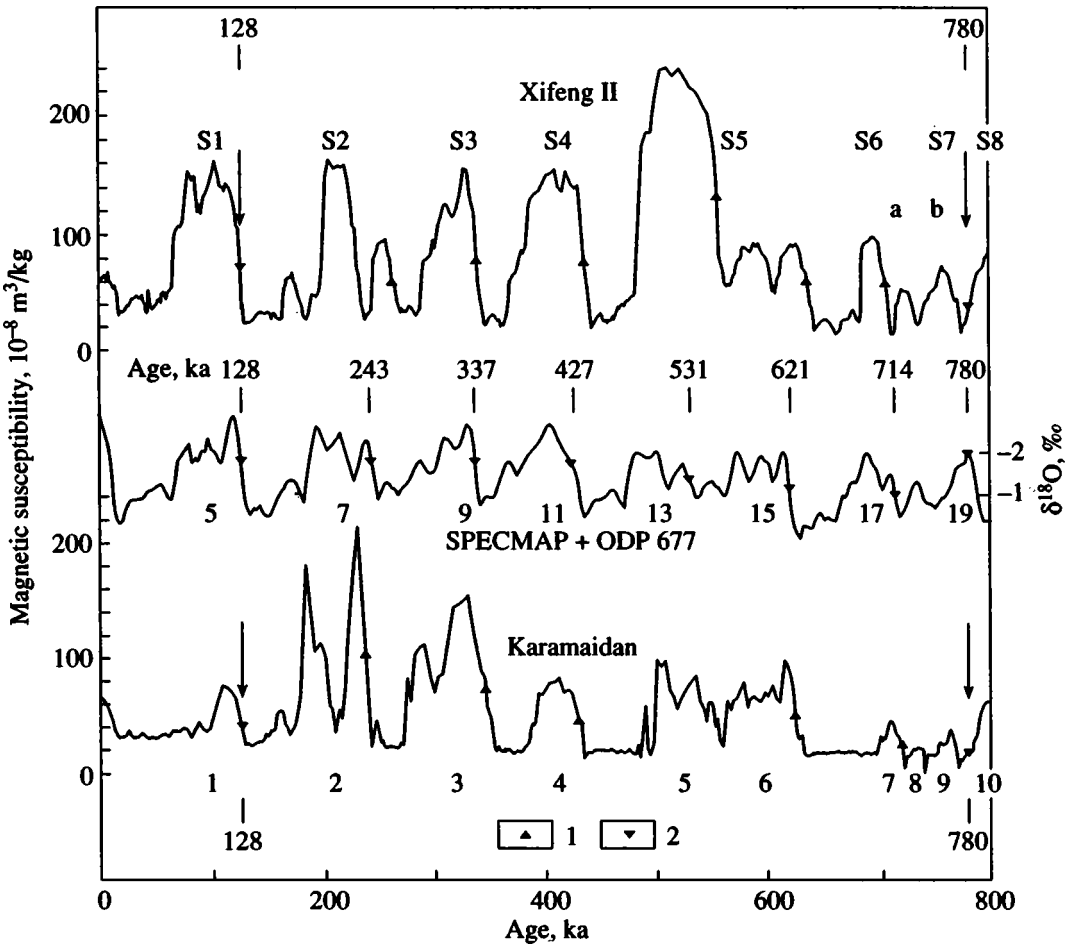

Fig. 5. Correlation of magnetic susceptibility curves (κ) for the loess–paleosol sections in the Karamaidan (Tajikistan) and Xifeng (China) with the oxygen isotope curve based on SPECMAP data and ODP Site 677. Two control levels (128 ka and the Brunhes–Matuyama reversal at 780 ka) are indicated by arrows (after Shackleton *et al.*, 1995): (1) boundary between loess bed and pedocomplex; (2) reference age levels.

world. For instance, paleoclimatic records near the Pleniglacial–Holocene boundary are delayed in the inner continental areas of the Middle East for three or four thousand years relative to similar events in coastal areas of the Middle and Near East regions (Roberts and Wright, 1993). Such a delay in climatic records cannot be excluded for some Pleistocene phases of the geological history of Central Asia.

PALEOCLIMATIC EVENTS AND EVOLUTION OF PALEOENVIRONMENT

The general trend of paleoclimatic changes in Central Asia during the late Pliocene–late Pleistocene time consists in transition from the relatively wet and hot conditions to arid and warm (or cool) environments, as is evident from paleontological records. For example, the late Pliocene fauna included some savanna inhabitants, along with dominant animals of open landscapes and the subordinate taxa of forest biotopes (Vangengeim *et al.*, 1988). The late Pliocene spectrum of palynomorphs contains relic elements of the American–East Asian vegetation (*Ostrya, Carya, Pterocarya*, and *Engelhardtia*) associated with highly diverse mesophilic palynoflora (Pakhomov *et al.*, 1980). In the late Pleistocene, the climate of Central Asia became more arid, approaching the continental type. The late Pleistocene fauna of piedmont zones was dominated by steppe forms. Some cold-resistant elements, such as *Coelodonta antiquitatis*, occurred as well. In the late Pleistocene, the distribution area of forest vegetation was reduced, whereas steppe and forest–steppe landscapes expanded.

The gradual transition to the arid continental climate is manifested in the structure of subaerial sedimentary sequences. The loess-forming process began in Central Asia 2.5–2.0 Ma ago under an arid climatic environment. The loess–paleosol succession exhibits alternating loess and paleosol horizons accumulated during arid and humid climatic phases, respectively. In the last period, from 0.8 Ma till present time, the alternation patterns of wet and arid phases were similar to those

Table 1. The loess–paleosol scale calibrated on the basis of the oxygen isotope stages. The age boundaries of the oxygen isotope stages are given in accordance with the SPECMAP curve and data on ODP Site 677 (Bassinot *et al.*, 1994). M/B denotes the Matuyama–Brunhes magnetic reversal.

Oxygen isotope stages	Age, ka	Loess horizons (L) and pedocomplexes (PC)
	11	
2		$L1_1$
	24	
3		$L1_{1-2}$
	57	
4		$L1_2$
	71	
5		PC1
	127	
6		L2
	186	
7		PC2
	242	
8		L3
	301	
9		PC3
	334	
10		L4
	364	
11		PC4
	427	
12		L5
	474	
13		PC5
	528	
14		L6
	568	
15		PC6
	621	
16		L7
	659	
17		PC7
	712	
18.2		L8
	724	
18.3		PC8
	741	
18.4		L9
	760	
M/B 19		PC9
	787	
20		L10 M/B
	806	
21		PC10
	860	

recorded in the oxygen isotope curve (Fig. 5). The periods of the piedmont soil formation differed in the moisture and temperature regimes.

The evolution of the paleoclimate and paleolandscapes since the time of PC10 formation is briefly characterized below. In this case, we use the results of paleosol studies and the data of spore–pollen analysis obtained from the loess–paleosol sequences by M.M. Pakhomov, A.A. Baiguzina, N.S. Bolikhovskaya, L.N. Ershova, and A.N. Simakova. The PC10 marks, in general, the predominance of forest–steppe landscapes. The forest vegetation expanded only in the optimal climatic phase corresponding to the lower horizon $PC10_2$. The formation time of PC9 and PC8 characterizes the expansion of steppe areas in the forest–steppe zone. Humidity parameters of these epochs were lower than they are now. The PC7 is formed in the forest–steppe landscape under conditions of increased precipitation, as compared to that of the periods of the PC9 and PC8 formation. The PC6 and PC5 are indicators of steppe and forest–steppe landscapes, where forest areas expanded during the climatic optima with relatively humid conditions. The developed part of PC4 profile is related in origin to the forest vegetation and has an elevated moisture content. Distinct illuvial–carbonate horizons of all lower pedocomplexes, including PC4, may be evidence of climate moistening due to precipitation during the warm seasons. The PC3 represents the soils of the steppe and forest–steppe landscapes, and the lower soil horizon $PC3_2$ reflects a higher moisture content than at the formation time of the upper horizon $PC3_1$. Forest–steppe and steppe landscapes were predominant during the formation of PC2 and PC1, when the influence of forest–steppe and forest ecosystems increased at the formation time of the lower soil horizons $PC2_2$ and $PC1_3$. The humidity and temperature fluctuated during the last period approximately in the same ranges, which are characteristic of the main subtypes of recent brown soils. Warm seasons of the period were arid.

The loess-forming epochs were associated with a considerable expansion of desert and desert–steppe landscapes in the piedmont zone. Steppes and related sierozem soils developed during the transitional periods from arid to humid climatic phases and vice versa. The growing aridity of the desert–steppe landscapes is recorded in the increased carbonate content of the respective loess horizons.

The loess sedimentation dependent on eolian processes was controlled by the air-mass dynamics. Paleoclimatic conditions in Central Asia were influenced by western air masses arriving from the North Atlantic and East Mediterranean, and also by the western branch of the Siberian anticyclone and Arctic air masses. Effect of continental air masses controlled by the Siberian anticyclone and Arctic winds prevailed during the glacials, whereas the western atmospheric flows were dominant during the interglacials. The inferable model of atmospheric processes shows that the Siberian (Siberian–Mongolian) anticyclone was most effective after the tectonic uplift of Tibet (Ruddiman and Kutzbach, 1991), which was at its peak (not less than 3000 m) in the late Pliocene–Pleistocene (Li Jiun *et al.*, 1995). To the west of Tibet, the simultaneous uplift was experienced by the Himalayas, Hindu Kush, Pamirs, and Tien Shan (Gansser, 1964; Burbank and Raynolds, 1984; Owen, 1989, Chediya, 1986). The mountain chains of the Himalayas and Hindu Kush blocked the northward movement of wet monsoons from the Indian Ocean. In the Pliocene–early Pleistocene, the precipitation was sufficient for forests to be widespread in the piedmont zone of the Pamirs and southern Tien Shan, but by the

Table 2. Sedimentation rates of loess for the last 0.8 Ma, the Darai Kalon section

Loess (L) horizon	Thickness of the horizon, m	Subsidence correction, %	Reconstructed thickness, m	Time range, thousand years	Sedimentation rate, mm/year
L1	18.65	–	–	60	0.31
L2	17.3	7	18.5	59	0.31
L3	8.1	10	8.9	59	0.15
L4	8.6	10	9.3	30	0.31
L5	11.4	10	12.5	47	0.27
L6	3.8	10	4.2	40	0.11
L7	10.8	10	11.9	38	0.31
L8	2.2	10	2.4	12	0.20
L9	3.9	10	4.3	19	0.23
L10	5.0	10	5.5	19	0.29

end of the Pleistocene, the distribution area of forests was noticeably reduced. To the north of the Hindu Kush, the precipitation during warm seasons and influence of southern monsoon appear to be completely ceased approximately at the end of the mid-Pleistocene, as it is recorded in the changes of soil profiles at the transition from PC4 to PC3 and younger pedocomplexes buried in the loess cover of southern Tajikistan.

Figure 6 shows the dynamics of loess formation during the last 0.8 Ma, which is established on the basis of sections studied in the southern Tajikistan sections. The Darai Kalon section (DRK in the Fig. 6) recorded the most stable development of subaerial sedimentation. Table 2 presents the estimated rates of dust accumulation in the Darai Kalon section. To reconstruct the initial thickness of loess horizons below 25–30 m, their measured thicknesses was reduced by 10% for the compaction (Kriger *et al.*, 1992). Table 2 shows as well that the dust accumulation rate varied in the range of 0.2–0.3 mm/year during the period under consideration. The minimal rate (0.11 mm/year) is characteristic of the loess horizon L6 and the maximal one (0.31 mm/year) of the horizons L1, L2, L4, and L7. During the soil formation, the estimated rate of dust accumulation varied in the range of 0.04–0.11 mm/year, if we accept the sedimentational model of paleosol development, i.e., the upward growth of a soil horizon owing to dust accumulation at the surface. However, this model cannot adequately explain the development of soil horizons, because it underestimates the fact that the soils develop reworking the underlying loess and increase in thickness incorporating the reworked mate-

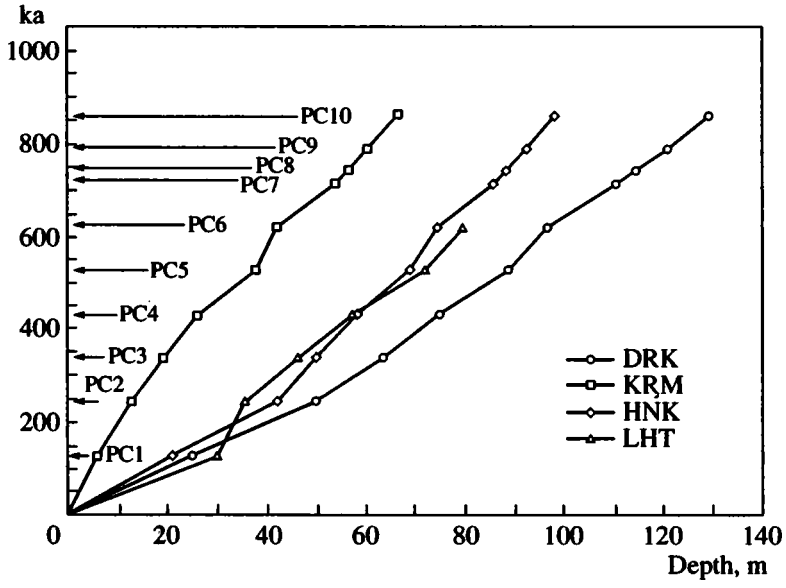

Fig. 6. Dynamics of loess accumulation in the southern Tajikistan; sections: (KRM) Karamaidan; (LHT) Lakhuti; (HNK) Khonako; (DRK) Darai Kalon. Arrows indicate lower age limits of pedocomplexes PC1, PC2, PC3, etc.

rial. In other words, the growing thickness of a soil horizon is under the control of two factors: the dust sedimentation and the partial inclusion of the bed rocks. Thus, we can reasonably suggest that, during the soil formation, the dust sedimentation rate was actually somewhat lower than that estimated above.

CONCLUSION

Recent investigations of loess–paleosol sequences in Central Asia showed that the thermoluminescence dating is most effective for the loess deposited during the last 100–130 thousand years. The TL ages of older loess horizons appear to be underestimated and this effect increases downsection. The controlling TL determinations at the level of the Matuyama–Brunhes paleomagnetic reversal (780 ka) yielded an apparently younger age of the deposits.

The detailed measurements of the magnetic susceptibility (κ) revealed magnetic properties of individual paleosols, which are used as preference parameters for the regional correlation of loess–paleosol sections. The correlation with the oxygen isotope scale and the usage of age markers (PC1 and the level of the Matuyama–Brunhes reversal) substantiated the chronological calibration of the loess–paleosol scale for the last 800 ka.

The loess-forming processes were affected by climatic fluctuations and air-mass dynamics. The loess accumulation was most intense during the glacials, when Central Asia was under the influence of the western periphery of the Siberian anticyclone and Arctic air masses. During the interglacials, there was the dominant influence of the western air mass and southern monsoon, which was active until the second half of the mid-Pleistocene.

Changes in soil profiles, which are recorded in PC4 and overlying pedocomplexes formed in the second half of the middle Pleistocene, can be interpreted as a result of a significant decrease in precipitation during the warm seasons caused by penetration of past southern monsoons. The rising mountain chains of the Himalayas represented an insuperable barrier for the monsoon advancing northward.

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Age of Upper Paleozoic Deposits in Eastern Margin of the Tunguska Syneclise: Implications from Palynological Data

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Abstract—Spore-and-pollen analysis of the Lapcha and Botuoba formations revealed four successive palynological assemblages. On the basis of the relative abundance of taxa and of the taxonomic composition of the assemblages, the Lapcha and Botuoba formations are attributed to the Upper Carboniferous and Lower Permian, respectively. The revealed palynofloras are correlated with the Upper Carboniferous and Lower Permian palynological zonation in the Kuznetsk coal basin.

Key words: *biostratigraphy, palynology, Carboniferous, Permian*

INTRODUCTION

Upper Paleozoic terrigenous deposits in the eastern margin of the Tunguska syneclise are intermediate collectors of diamonds and attract the attention of many researchers. Available data on stratigraphy, lithology, and sedimentation environments of these deposits are quite representative and have already been considered in a series of publications (Vashchenko and Fainshtein, 1985; *Geologiya ...*, 1986; Saltykov *et al.*, 1991; and some others).

Within the region under consideration, the Upper Paleozoic sequence varies in thickness from 0 to 250 m and is subdivided into the Lapcha (C_2^1), Botuoba (C_2^2–C_3), Akhtaranda (P_1), Borulloi (P_2), and Dergala (P_2) formations, in accordance with the reference legend approved by the Scientific Editorial Council of the USSR Ministry of Geology in 1989. However, the age of the units is still in dispute (Chernaya, 1996). Leaving aside all viewpoints concerning this problem, I should mention that even experts in palynology interpret the same subdivisions as varying in age from the Middle Carboniferous to the Early Permian (Mikhailova, 1991; Chernaya, 1991). Consequently, it would be purposeful to consider once again the results of palynological study of 31 rock samples donated by S.F. Adel'shina (Chernyshevskii Expedition, PGO "Yakutgeologiya") to the All-Russia Research Institute of Geology.

The samples have been collected in the Chernyshevskii and Akhtaranda areas (left-hand side of the Vilyui River) within the eastern margin of the Tunguska syneclise (80–120 km to the west and northwest off the town of Mirnyi, Fig. 1). They characterize the Lapcha and Botuoba formation. Adel'shina preliminary considered both formations as corresponding in age to the Middle–Late Carboniferous (C_2^1 lp; C_2^2 bt_1; C_3bt_2).

DESCRIPTION OF PALYNOFLORAS

The results of my investigation allow me to distinguish four palynofloras characterizing deposits of the Lapcha and Botuoba formations. Their taxonomic composition and abundance rates (%) of individual species in some pollen-and-spore fractions from the palynofloras are presented in Table 1.

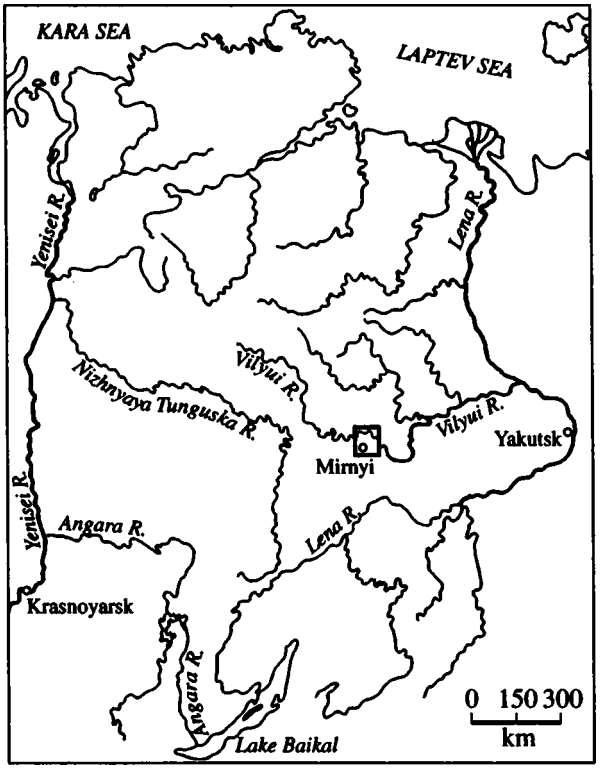


Fig. 1. Geographic position of the study area.

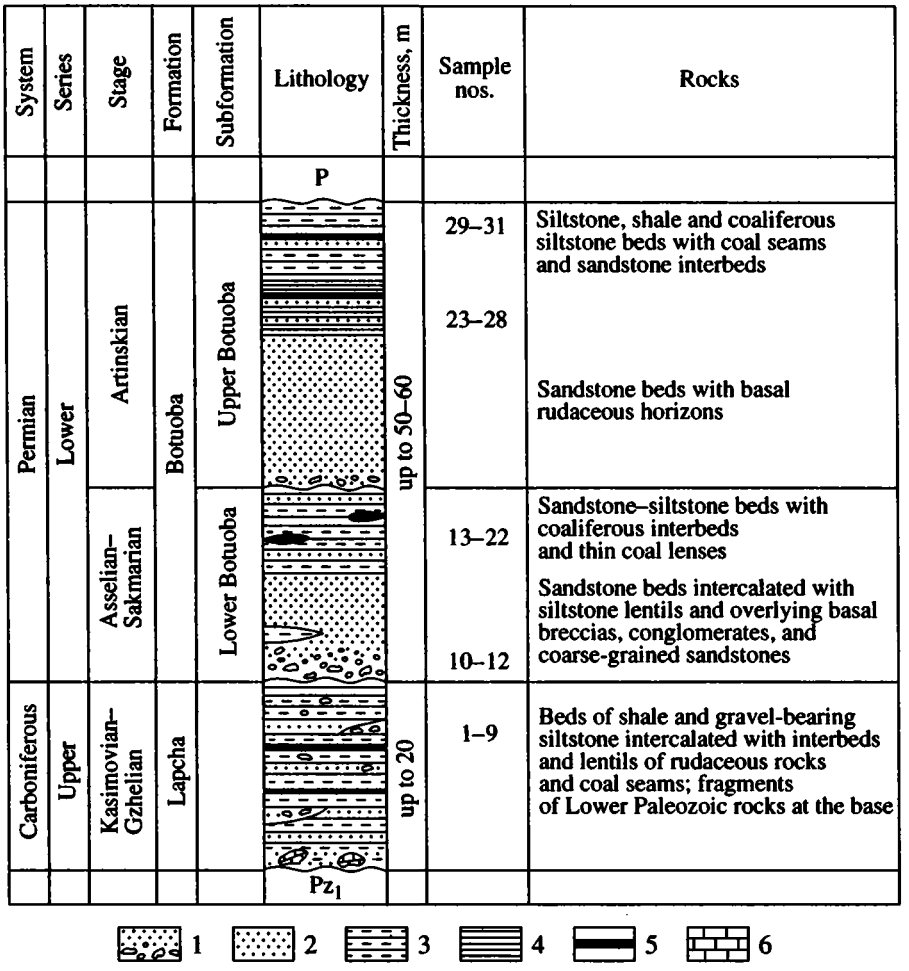


Fig. 2. Composite lithology and stratigraphy of the Lapcha and Botuoba formations in the western margin of the Tunguska syneclise: (1) conglomerate, (2) sandstone, (3) siltstone, (4) shale, (5) coal, (6) limestone. Sampled localities: nos. 1–9 from wells 230/990, 63.8/28, 61/34, and 74.2/170; 10–12 from Well 140/895; nos. 13–22 from wells 140/895, 76/395, 120/570 and from test pits 569/3 and 530/3; nos. 23–31 from wells 120/570 and 230/910.

The first palynoflora is detected in the Lapcha Formation (Fig. 2). Nine studied samples represent siltstone from wells 230/990 (depth levels of 230.0 and 227.4 m), 63.8/28 (depth levels of 42.0, 32.0 and 26.0 m) and 61/34 (depth levels of 109.2 and 108.2 m), and also coaliferous siltstones from Well 74.2/170 (depth levels of 141.9 and 138.5 m). Palynological preparations macerated from the samples are rather rich in palynomorphs of a good preservation degree in general. Macerations include rather abundant fragments of plant tissue and rare fragmented cuticles. Almost all samples yielded singular fungal spores *Sporonites unionus* (Horst) Dyb. et Jach. and sporadic remains of alga *Pilla.*

Spore-and-pollen assemblages from the Lapcha Formation persistently include a noticeable amount (13.0–44.5%) of spores *Psilohymena psiloptera* (Lub.) Hart et Harr. and *P. mirabilis* (Lub.) Hart et Harr in association with less abundant spores *Retusotriletes nigritellus* (Lub.) Fost. (1.0–6.5%) and monosaccate pollen of *Florinites* (2.0–18.0%) and *Plicatipollenites* (1.0–10.0%). Species *Cyclobaculisporites trichacanthus* Lub., *Horriditriletes trichacanthus* (Lub.) Oshurk., *Leiotriletes pyramidalis* (Lub.) All., and *Acanthotriletes rigidispinosus* (Lub.) Isch. are completely missing from some assemblages being as abundant as 7.0% in others. Abundance of *Cordaitina* pollen varies from 0 to 23.5%, and that of *Potonieisporites* from 0 to 6.0%. In general, palynoflora from the Lapcha Formation shows a distinct preponderance of spores (65.0–90.0%) over pollen grains (10.0–35.0%) that is characteristic of Carboniferous deposits. Pollen of *Crucisaccites* (0–2.0%), *Bascanisporites* (0–0.5%), *Taenisporites* (0–1.0%), *Densipollenites* (0–2.0%), *Platysaccus* (0–3.0%), *Protohaploxypinus* (0–4.0%), *Gardenosporites* (0–2.5%), costate pollen of *Striatosaccites* (0–3.0%), and bisaccate pollen of *Disaccites* (0–2.5%) can be used to circumstantiate the age and to attribute the host deposits to the Upper Carboniferous (Plate I, nos. 1–7).

The second palynoflora characterizes the lower beds of the lower Botuoba Subformation (Fig. 2). Palynological assemblages of three siltstone samples from Well 140/895 (depth levels of 141.8, 140.0, and 139.5 m) are

Table 1. Species composition and percentage of miospores in palynological spectra from the Lapcha and Botuoba formations

Palynoflora no.	I		II		III				IV	
Well no.	230/990	74.2/170	140/895	140/895	140/895	76/395	76/395	120/570	120/570	230/910
depth, m	227.4	138.5	141.8	139.5	78.0	108.0	97.0	210.4	163.4	131.0
Species composition										
Calamospora microrugosa (Ibr.) S., W. et B.	0.5	5.0	2.0	–	2.0	–	1.0	–	–	–
Punctatisporites glabratus Lub.	–	11.0	–	2.0	–	1.0	–	–	–	–
P. platirugosus (Waltz) Sulliv.	4.5	4.5	–	–	2.0	–	–	1.0	1.0	–
P. gretensis Balme et Henn.	–	3.5	–	–	–	–	–	–	–	–
Leiotriletes extensus (Lub.) Siv.	1.0	0.5	4.0	1.0	2.0	–	–	3.0	2.0	2.0
L. pyramidalis (Lub.) Medv.	–	1.0	–	–	6.0	–	2.0	–	1.0	–
L. trimodius (Andr.) Pashk.	–	–	3.0	–	–	–	–	–	–	–
Retusotriletes nigritellus (Lub.) Fost.	4.5	6.5	1.0	5.0	1.0	4.0	3.0	1.0	7.0	7.0
Cyclogranisporites sp.	–	–	–	–	–	–	–	–	–	1.0
Granulatisporites pulvigerus Lub.	–	–	1.0	–	–	1.0	–	–	1.0	–
G. parvus (Ibr.) Pot. et Kr.	–	–	–	–	1.0	–	2.0	2.0	–	–
G. piroformis Loose	2.0	3.5	–	1.0	–	–	–	–	–	2.0
Tuberculispora verruculosa (Drjag.) Oshurk.	–	–	–	1.0	–	–	–	–	–	–
Lophotriletes resistens (Lub.) Medv.	0.5	1.0	1.0	–	–	–	2.0	–	1.0	–
Verrucosisporites sp.	–	–	1.0	–	1.0	1.0	2.0	–	–	5.0
Cyclobaculisporites trichacanthus Lub.	7.0	2.5	1.0	–	7.0	5.0	2.0	20.0	5.0	–
C. muticus Drjag.	–	–	–	–	–	–	–	–	–	8.0
Horriditriletes trichacanthus (Lub.) Oshurk.	0.5	–	4.0	1.0	6.0	4.0	1.0	10.0	3.0	–
Microbaculispora tentula Tiwari	–	2.0	1.0	–	–	–	2.0	1.0	–	–
Cyclopileatispora gibberula (Lub.) Oshurk.	–	–	6.0	8.0	5.0	4.0	11.0	–	–	–
Acanthotriletes rigidispinosus (Lub.) Isch.	2.0	6.0	1.0	2.0	1.0	–	1.0	–	1.0	–
A. facerus (Andr.) Siv.	–	–	–	–	2.0	4.0	1.0	–	–	–
A. instabilis (Drjag.) Oshurk.	–	–	–	–	–	–	–	–	4.0	–
Apiculatisporis hispidus (Andr.) Drjag.	3.0	3.5	–	–	–	–	–	–	4.0	–
A. curvispinus (Lub.) Oshurk.	4.0	–	3.0	–	2.0	2.0	1.0	–	–	–
Iugisporis parvispinus (Lub.) Oshurk.	–	–	–	–	2.0	–	–	–	–	–
Convolutispora sp.	1.5	–	–	–	–	–	–	–	–	–
Neoraistrickia obtusosaetosa (Lub.) Oshurk.	–	–	–	–	1.0	–	–	–	4.0	2.0
Lophozonotriletes depressus Drjag.	2.5	–	–	–	–	–	–	–	–	–
Knoxisporites sp.	–	2.0	–	–	–	–	–	–	–	–
Krauselisporites sp.	–	–	–	–	–	4.0	1.0	–	–	–
Psilohymena mirabilis (Lub.) Hart et Harr.	4.5	3.5	1.0	1.0	–	–	–	2.0	–	–
P. psiloptera (Lub.) Hart et Harr.	40.0	33.0	17.0	23.0	2.0	12.0	2.0	5.0	5.0	5.0
Lycospora rotunda (Bharad.) Somers *et al.*	4.5	5.0	–	–	–	–	–	–	3.0	2.0
Spores in total	78.0	89.0	47.0	45.0	43.0	42.0	35.0	45.0	42.0	34.0
Potonieisporites novicus Bhard.	1.5	–	2.0	4.0	1.0	–	2.0	3.0	1.0	2.0
Florinites visendus (Ibr.) S., W. et B.	3.5	4.5	3.0	4.0	3.0	3.0	3.0	2.0	2.0	–
F. pumicosus (Ibr.) S., W. et B.	–	–	5.0	1.0	–	–	–	–	–	–

Table 1. (Contd.)

Palynoflora no.	I		II		III				IV	
Well no.	230/990	74.2/170	140/895	140/895	140/895	76/395	76/395	120/570	120/570	230/910
depth, m	227.4	138.5	141.8	139.5	78.0	108.0	97.0	210.4	163.4	131.0
F. macropterus (Lub.) Dib.	–	–	–	–	–	–	1.0	–	–	–
F. grandis (Lub.) Dibn.	–	–	2.0	5.0	–	–	–	–	–	–
F. luberae Samoil.	–	–	3.0	1.0	–	–	–	1.0	–	1.0
Densipollenites sp.	–	–	1.0	–	–	–	–	–	–	–
Cordaitina rugulifer (Lub.) Samoil.	12.0	–	8.0	20.0	18.0	3.0	8.0	9.0	6.0	5.0
C. rotata (Lub.) Medv.	–	–	1.0	–	4.0	12.0	6.0	10.0	12.0	10.0
C. uralensis (Lub.) Samoil.	2.0	–	–	4.0	13.0	6.0	9.0	11.0	6.0	5.0
C. abutiloides (Andr.) Dibn.	–	–	2.0	–	1.0	2.0	7.0	3.0	2.0	5.0
C. angustilimbata (Lub.) Varjuch.	–	–	2.0	–	1.0	–	–	5.0	–	1.0
C. convallata (Lub.) Samoil.	–	–	–	–	–	–	1.0	–	–	–
C. punctata (Lub.) Hart	–	–	–	–	–	–	–	–	–	3.0
Crucisaccites ornatus (Samoil.) Dibn.	–	0.5	3.0	4.0	5.0	22.0	8.0	2.0	4.0	2.0
Bascanisporites verus (Sadk.) Dibn.	–	0.5	–	–	–	–	–	–	1.0	4.0
B. stipticus (Lub.) Oshurk.	0.5	–	–	1.0	5.0	–	2.0	–	2.0	12.0
B. varians (Sadk.) Dibn.	–	–	–	–	1.0	–	–	–	6.0	2.0
Plicatipollenites gondwanensis (B. et H.) Lele	2.5	2.0	12.0	5.0	1.0	5.0	13.0	3.0	5.0	2.0
P. indicus Lele	–	–	–	–	1.0	–	–	–	–	–
P. densus Sriv.	–	–	–	1.0	–	–	–	–	–	–
Luberisaccites subrotatus (Lub.) Dibn.	–	–	4.0	1.0	–	2.0	–	–	–	1.0
L. geminus (Andr.) Dibn.	–	–	1.0	–	–	–	–	–	–	–
Disaccites	–	0.5	2.0	2.0	–	3.0	1.5	1.0	–	–
Illinites sp.	–	–	–	–	–	–	–	1.0	–	–
Limitisporites moersensis (Grebe) Klaus	–	–	–	–	–	–	–	–	1.0	–
L. monstruosus (Lub.) Hart	–	–	–	–	–	–	0.5	–	–	–
Platysaccus sp.	–	–	0.5	1.0	1.0	–	–	–	–	1.0
Protohaploxypinus varius (Lesch.) Oshurk.	–	–	–	2.0	–	–	–	–	3.0	–
P. perfectus (Naum. in K.-M.) Samoil.	–	0.5	–	–	1.0	–	–	1.0	2.0	1.0
Strotersporites marginalis Kl.	–	–	–	–	–	–	2.0	–	–	–
Vittatina striata (Lub.) Samoil.	–	–	–	–	–	–	–	–	–	1.0
Marsupipollenites subrotatus (Lub.) Oshurk.	–	–	–	–	–	–	1.0	–	–	1.0
M. retroflexus Lub.	–	1.0	–	–	–	–	–	–	–	1.0
Entylissa caperata (Lub.) Pot. et Kr.	–	–	1.0	–	1.0	–	–	1.0	2.0	–
Pollen in total	**22.0**	**10.0**	**53.0**	**55.0**	**57.0**	**58.0**	**65.0**	**55.0**	**55.0**	**60.0**
Azonaletes levis	–	1.0	–	–	1.0	–	–	–	3.0	6.0
Total amount of counted miospores	200	200	200	200	200	200	200	200	200	200

similar in general to those from the Lapcha Formation. They again include the rather abundant spores *Psilohymena psiloptera* (Lub.) Hart et Harr. (17.0–30.0%), in association with *Cyclobaculisporites trichacanthus* Lub., *Horriditriletes trichacanthus* (Lub.) Oshurk., and *Cyclopileatispora gibberula* (Lub.) Oshurk. (9.0–12.0%), and also with *Leiotriletes pyramidalis* (Lub.) All., *Acanthotriletes rigidispinosus* (Lub.) Isch. (1.0–5.0%) and *Retusotriletes nigritellus* (Lub.) Fost. (1.0–6.0%). Abundance of monosaccate pollen of *Florinites* (7.0–14.0%) and *Plicatipollenites* also remains more or less at the former level. However, in contrast to the Lapcha assemblages, *Cordaitina* pollen becomes more abundant (15.0–29.0%) and shows a higher species diversity (Table 1) in the samples under consideration. Almost equal abundance rates of spores (47–57%) and pollen (43–53%) in the samples also distinguish the second palynoflora from that of the Lapcha Formation, where spores are distinctly predominant. The greater abundance of pollen in general and higher species diversity of *Cordaitina* pollen point to the basal Lower Permian level of palynoflora in question (Plate I, nos. 8–15).

Table 1. Species composition and percentage of miospores in palynological spectra from the Lapcha and Botuoba formations

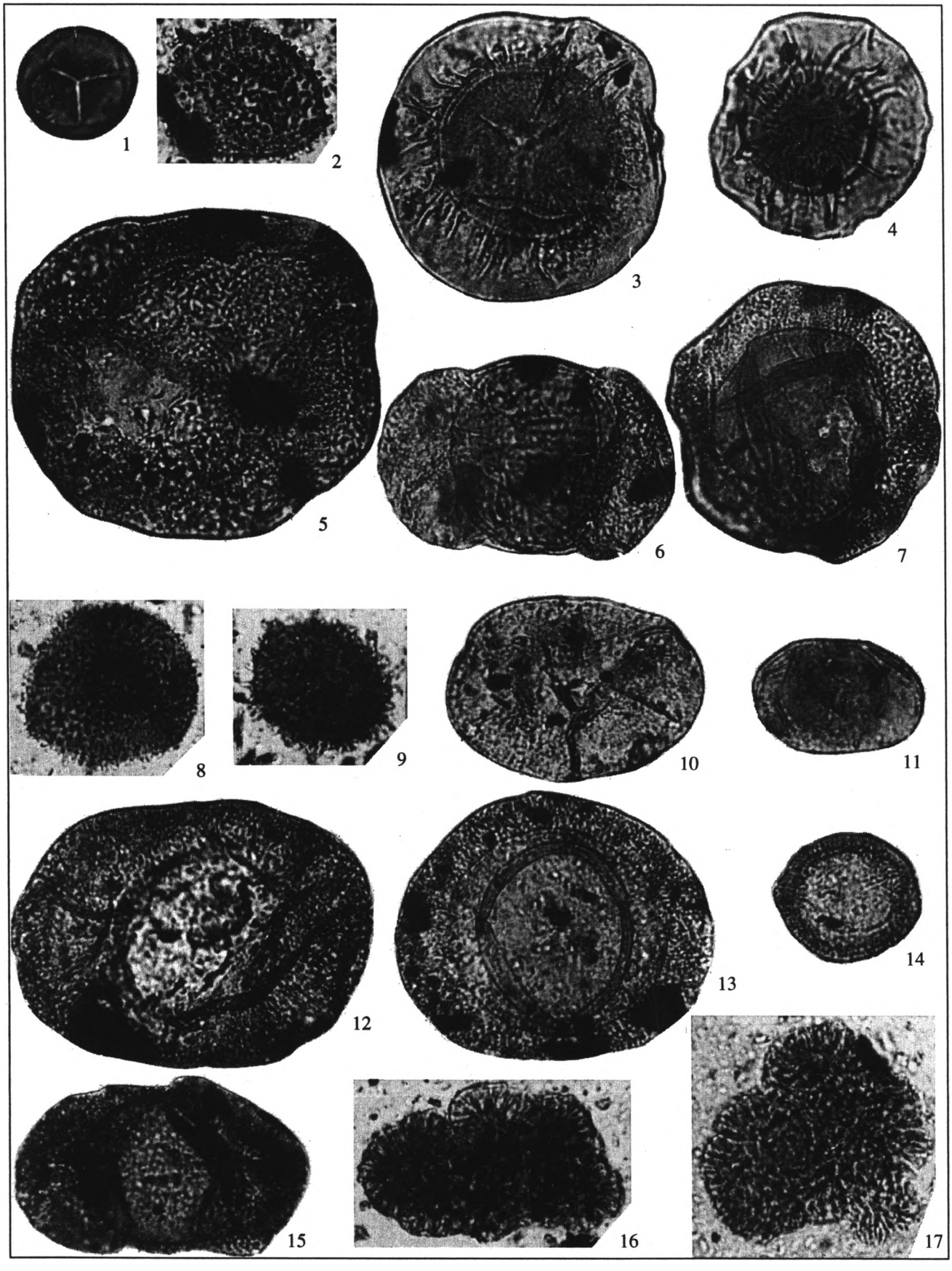
Palynoflora no.	I		II		III				IV	
Well no.	230/990	74.2/170	140/895	140/895	140/895	76/395	76/395	120/570	120/570	230/910
depth, m	227.4	138.5	141.8	139.5	78.0	108.0	97.0	210.4	163.4	131.0
Species composition										
Calamospora microrugosa (Ibr.) S., W. et B.	0.5	5.0	2.0	–	2.0	–	1.0	–	–	–
Punctatisporites glabratus Lub.	–	11.0	–	2.0	–	1.0	–	–	–	–
P. platirugosus (Waltz) Sulliv.	4.5	4.5	–	–	2.0	–	–	1.0	1.0	–
P. gretensis Balme et Henn.	–	3.5	–	–	–	–	–	–	–	–
Leiotriletes extensus (Lub.) Siv.	1.0	0.5	4.0	1.0	2.0	–	–	3.0	2.0	2.0
L. pyramidalis (Lub.) Medv.	–	1.0	–	–	6.0	–	2.0	–	1.0	–
L. trimodius (Andr.) Pashk.	–	–	3.0	–	–	–	–	–	–	–
Retusotriletes nigritellus (Lub.) Fost.	4.5	6.5	1.0	5.0	1.0	4.0	3.0	1.0	7.0	7.0
Cyclogranisporites sp.	–	–	–	–	–	–	–	–	–	1.0
Granulatisporites pulvigerus Lub.	–	–	1.0	–	–	1.0	–	–	1.0	–
G. parvus (Ibr.) Pot. et Kr.	–	–	–	–	1.0	–	2.0	2.0	–	–
G. piroformis Loose	2.0	3.5	–	1.0	–	–	–	–	–	2.0
Tuberculispora verruculosa (Drjag.) Oshurk.	–	–	–	1.0	–	–	–	–	–	–
Lophotriletes resistens (Lub.) Medv.	0.5	1.0	1.0	–	–	–	2.0	–	1.0	–
Verrucosisporites sp.	–	–	1.0	–	1.0	1.0	2.0	–	–	5.0
Cyclobaculisporites trichacanthus Lub.	7.0	2.5	1.0	–	7.0	5.0	2.0	20.0	5.0	–
C. muticus Drjag.	–	–	–	–	–	–	–	–	–	8.0
Horriditriletes trichacanthus (Lub.) Oshurk.	0.5	–	4.0	1.0	6.0	4.0	1.0	10.0	3.0	–
Microbaculispora tentula Tiwari	–	2.0	1.0	–	–	–	2.0	1.0	–	–
Cyclopileatispora gibberula (Lub.) Oshurk.	–	–	6.0	8.0	5.0	4.0	11.0	–	–	–
Acanthotriletes rigidispinosus (Lub.) Isch.	2.0	6.0	1.0	2.0	1.0	–	1.0	–	1.0	–
A. facerus (Andr.) Siv.	–	–	–	–	2.0	4.0	1.0	–	–	–
A. instabilis (Drjag.) Oshurk.	–	–	–	–	–	–	–	–	4.0	–
Apiculatisporis hispidus (Andr.) Drjag.	3.0	3.5	–	–	–	–	–	–	4.0	–
A. curvispinus (Lub.) Oshurk.	4.0	–	3.0	–	2.0	2.0	1.0	–	–	–
Iugisporis parvispinus (Lub.) Oshurk.	–	–	–	–	2.0	–	–	–	–	–
Convolutispora sp.	1.5	–	–	–	–	–	–	–	–	–
Neoraistrickia obtusosaetosa (Lub.) Oshurk.	–	–	–	–	1.0	–	–	–	4.0	2.0
Lophozonotriletes depressus Drjag.	2.5	–	–	–	–	–	–	–	–	–
Knoxisporites sp.	–	2.0	–	–	–	–	–	–	–	–
Krauselisporites sp.	–	–	–	–	–	4.0	1.0	–	–	–
Psilohymena mirabilis (Lub.) Hart et Harr.	4.5	3.5	1.0	1.0	–	–	–	2.0	–	–
P. psiloptera (Lub.) Hart et Harr.	40.0	33.0	17.0	23.0	2.0	12.0	2.0	5.0	5.0	5.0
Lycospora rotunda (Bharad.) Somers *et al.*	4.5	5.0	–	–	–	–	–	–	3.0	2.0
Spores in total	78.0	89.0	47.0	45.0	43.0	42.0	35.0	45.0	42.0	34.0
Potonieisporites novicus Bhard.	1.5	–	2.0	4.0	1.0	–	2.0	3.0	1.0	2.0
Florinites visendus (Ibr.) S., W. et B.	3.5	4.5	3.0	4.0	3.0	3.0	3.0	2.0	2.0	–
F. pumicosus (Ibr.) S., W. et B.	–	–	5.0	1.0	–	–	–	–	–	–

Table 1. (Contd.)

Palynoflora no.	I		II		III				IV	
Well no.	230/990	74.2/170	140/895	140/895	140/895	76/395	76/395	120/570	120/570	230/910
depth, m	227.4	138.5	141.8	139.5	78.0	108.0	97.0	210.4	163.4	131.0
F. macropterus (Lub.) Dib.	–	–	–	–	–	–	1.0	–	–	–
F. grandis (Lub.) Dibn.	–	–	2.0	5.0	–	–	–	–	–	–
F. luberae Samoil.	–	–	3.0	1.0	–	–	–	1.0	–	1.0
Densipollenites sp.	–	–	1.0	–	–	–	–	–	–	–
Cordaitina rugulifer (Lub.) Samoil.	12.0	–	8.0	20.0	18.0	3.0	8.0	9.0	6.0	5.0
C. rotata (Lub.) Medv.	–	–	1.0	–	4.0	12.0	6.0	10.0	12.0	10.0
C. uralensis (Lub.) Samoil.	2.0	–	–	4.0	13.0	6.0	9.0	11.0	6.0	5.0
C. abutiloides (Andr.) Dibn.	–	–	2.0	–	1.0	2.0	7.0	3.0	2.0	5.0
C. angustilimbata (Lub.) Varjuch.	–	–	2.0	–	1.0	–	–	5.0	–	1.0
C. convallata (Lub.) Samoil.	–	–	–	–	–	–	1.0	–	–	–
C. punctata (Lub.) Hart	–	–	–	–	–	–	–	–	–	3.0
Crucisaccites ornatus (Samoil.) Dibn.	–	0.5	3.0	4.0	5.0	22.0	8.0	2.0	4.0	2.0
Bascanisporites verus (Sadk.) Dibn.	–	0.5	–	–	–	–	–	–	1.0	4.0
B. stipticus (Lub.) Oshurk.	0.5	–	–	1.0	5.0	–	2.0	–	2.0	12.0
B. varians (Sadk.) Dibn.	–	–	–	–	1.0	–	–	–	6.0	2.0
Plicatipollenites gondwanensis (B. et H.) Lele	2.5	2.0	12.0	5.0	1.0	5.0	13.0	3.0	5.0	2.0
P. indicus Lele	–	–	–	–	1.0	–	–	–	–	–
P. densus Sriv.	–	–	–	1.0	–	–	–	–	–	–
Luberisaccites subrotatus (Lub.) Dibn.	–	–	4.0	1.0	–	2.0	–	–	–	1.0
L. geminus (Andr.) Dibn.	–	–	1.0	–	–	–	–	–	–	–
Disaccites	–	0.5	2.0	2.0	–	3.0	1.5	1.0	–	–
Illinites sp.	–	–	–	–	–	–	–	1.0	–	–
Limitisporites moersensis (Grebe) Klaus	–	–	–	–	–	–	–	–	1.0	–
L. monstruosus (Lub.) Hart	–	–	–	–	–	–	0.5	–	–	–
Platysaccus sp.	–	–	0.5	1.0	1.0	–	–	–	–	1.0
Protohaploxypinus varius (Lesch.) Oshurk.	–	–	–	2.0	–	–	–	–	3.0	–
P. perfectus (Naum. in K.-M.) Samoil.	–	0.5	–	–	1.0	–	–	1.0	2.0	1.0
Strotersporites marginalis Kl.	–	–	–	–	–	–	2.0	–	–	–
Vittatina striata (Lub.) Samoil.	–	–	–	–	–	–	–	–	–	1.0
Marsupipollenites subrotatus (Lub.) Oshurk.	–	–	–	–	–	–	1.0	–	–	1.0
M. retroflexus Lub.	–	1.0	–	–	–	–	–	–	–	1.0
Entylissa caperata (Lub.) Pot. et Kr.	–	–	1.0	–	1.0	–	–	1.0	2.0	–
Pollen in total	**22.0**	**10.0**	**53.0**	**55.0**	**57.0**	**58.0**	**65.0**	**55.0**	**55.0**	**60.0**
Azonaletes levis	–	1.0	–	–	1.0	–	–	–	3.0	6.0
Total amount of counted miospores	200	200	200	200	200	200	200	200	200	200

similar in general to those from the Lapcha Formation. They again include the rather abundant spores *Psilohymena psiloptera* (Lub.) Hart et Harr. (17.0–30.0%), in association with *Cyclobaculisporites trichacanthus* Lub., *Horriditriletes trichacanthus* (Lub.) Oshurk., and *Cyclopileatispora gibberula* (Lub.) Oshurk. (9.0–12.0%), and also with *Leiotriletes pyramidalis* (Lub.) All., *Acanthotriletes rigidispinosus* (Lub.) Isch. (1.0–5.0%) and *Retusotriletes nigritellus* (Lub.) Fost. (1.0–6.0%). Abundance of monosaccate pollen of *Florinites* (7.0–14.0%) and *Plicatipollenites* also remains more or less at the former level. However, in contrast to the Lapcha assemblages, *Cordaitina* pollen becomes more abundant (15.0–29.0%) and shows a higher species diversity (Table 1) in the samples under consideration. Almost equal abundance rates of spores (47–57%) and pollen (43–53%) in the samples also distinguish the second palynoflora from that of the Lapcha Formation, where spores are distinctly predominant. The greater abundance of pollen in general and higher species diversity of *Cordaitina* pollen point to the basal Lower Permian level of palynoflora in question (Plate I, nos. 8–15).

Plate I



Another peculiar feature of rock samples from the base of the lower Botuoba Subformation is a high abundance of remains of planktonic yellow–green botryococcalean algae (Plate I, nos. 16, 17).

The third palynoflora from the upper part of the lower Botuoba Subformation (Fig. 2) is macerated from ten rock samples recovered by wells 76/395 (depth levels of 108.0, 103.7, and 97.0 m), 120/570 (depth levels of 211.0 and 210.4 m), 140/895 (depth levels of 78.0 and 77.0 m), and also in test pits 569/3 (depth level of 4.0 m) and 530/3 (depth levels of 21.8 and 33.2 m). All samples, except two coaliferous rock varieties from Well 120/570, correspond in composition to siltstones. Spore-and-pollen assemblages from this part of the section persistently include the quite abundant *Cordaitina* pollen that associates sometimes with *Crucisaccites* pollen (12.0–42.0%), whose abundance may be as high as 22.0%. Contents somewhat increased at this level are characteristic of *Disaccites* (up to 3.0%), *Stratersporites* (up to 3.0%), and *Bascanisporites* (up to 6.0%) pollen. Proportions of *Psilohymena* spores (1.0–15.0%) are considerably lower, but spores *Retusotriletes nigritellus* (Lib.) Fost. (1.0–8.0%), *Leiotriletes pyramidalis* (Lub.) All., *Lophotriletes resistens* (Lub.) Medv., and *Acanthotriletes rigidispinosus* (Lub.) Isch. (0–7.0%) are still as abundant as before. Samples of coaliferous siltstone are distinct amid others because of a considerable abundance of spores *Cyclobaculisporites trichacanthus* Lub. and *Horriditriletes trichacanthus* (Lub.) Oshurk. (11.0–30.0%). Thus, the third palynoflora appears to be inherited from basal beds of the lower Botuoba Subformation, though it includes the new spore species *Krauselisporites* (0–4.0%), *Densoisporites* (0–18.0%), and *Vittatina* pollen (0–1.0%), all characteristic of Permian deposits. It is likely that the grown general proportion of pollen (up to 65%), the parallel decrease of spore abundance (35%), and a series of newcomers typical of Permian time (Plate II, nos. 1–14) confidently suggest the Early Permian age of the upper part of the lower Botuoba Subformation.

One more distinctive feature of this interval is the extremely low amount of algae remains in macerations and, in contrast, the permanent presence of acritarchs *Inderites compactus* (Lub.) Abr. et March., *I. spinereticulatus* Abr. et March., *Tessellaesphaera tessellata* Fost., *Leiosphaeridia* sp., and *Tasmanites* sp. (Plate II, nos. 13, 14), which are not abundant and associate with fungal spores *Sporonites*.

The peculiar planktonic microorganisms occurring in macerations from the lower and upper parts of the lower Botuoba Subformation, namely yellow-green botryococcalean algae typical of fresh-water basins (first case) and acritarchs dwelling in marine basins (second case), suggest that sedimentation environments changed from continental-lacustrine to shallow-marine, respectively. This conclusion is compatible with the Late Paleozoic paleogeography of the eastern margin of the Tunguska syneclise (Saltykov *et al.*, 1991).

The fourth palynoflora of the upper Botuoba Subformation (Fig. 2) is macerated from nine siltstone samples from wells 120/570 (depth levels of 163.4 and 159.5 m) and 230/910 (depth levels of 136.2, 133.9, 131.0, 129.2, 126.8, 126.6, and 124.0 m). All samples are moderately rich in dispersed miospores of a low preservation degree. Macerations usually include fragments of plant tissue and cuticles. Remains of acritarchs *Inderites, Tasmanites*, Leaeosphaerae, and fungal spores *Sporonites* are also persistent components. As compared to assemblages of the third palynoflora from the upper part of the lower Botuoba Subformation, palynomorphs of the fourth level are rich in *Cordaitina* pollen (16.0–41.0%) constantly associated with *Plicatipollenites* (2.0–9.0%), *Potonieisporites* (1.0–5.0%), and *Crucisaccites* (1.0–4.0%) pollen. Somewhat less abundant spore taxa are *Cyclobaculisporites trichacanthus* Lub., *Horriditriletes trichacanthus* (Lub.) Oshurk., *H. mamillosus* (Drjag.) Oshurk., (6.0–17.0%), *Lophotriletes resistens* (Lub.) Oshurk., *Acanthotriletes rigidispinosus* Isch. (0–3.0%), and *Psilohymena* sp. (2.0–11.0%). In contrast, the content of spores *Leiotriletes* sp. (2.0–14.0%), *Retusotriletes nigritellus* (Lub.) Fost. (2.0–15.0%), and *Lycospora* sp. (1.0–12.0%) is increased. The growing abundance in pollen assemblages is characteristic of *Bascanisporites verus* (Sadk.) Dibn., *B. varians* (Sadk.) Dibn., *B. stipticus* (Lub.) Oshurk., (4.0–12.0%), *Protohaploxypinus* sp. (0–5.0%), *Striatosaccites* (0–3.0%), and *Marsupipollenites* (0–6.0%). The taxonomic composition of spore-and-pollen assemblages in question is more diverse than before (Table 1) and includes species missing from underlying deposits: *Neoraistrickia* sp. (0–7.0%), *Cirratriradites gracilis* Drjag. (0–2.0%), *Laevigatosporites* sp. (0–1.0%), *Jugasporites delasaucei* (Pot. et Kr.) Lesch. (0–3.0%), *Alisporites* (0–1.0%), and

Plate I. Basic palynomorphs of palynofloras I and II

(1–7) Palynoflora I from the Lapcha Formation (C_3) (nos. 1, 4 from Well 74.2/170, depth levels of 138.5 and 141.9 m; nos. 2, 6 from Well 63.8/28, depth level of 42.0 m; nos. 3, 5, and 7 from Well 61/34, depth level of 108.2 m): (1) *Retusotriletes nigritellus* (Luber) Foster; (2) *Horriditriletes trichacanthus* (Luber) Oshurkova, comb. nov.; (3) *Psilohymena psiloptera* (Luber) Pashkevich; (4) *Psilohymena mirabilis* (Luber) Hart et Harrison; (6) *Protohaploxypinus hartii* Foster; (7) *Plicatipollenites indicus* Lele. (8–15) Palynoflora II from basal beds of the lower Botuoba Subformation (P_1^1) (nos. 8–17 from Well 140/895; nos. 10, 12–14, 17 from depth level of 139.5 m; nos. 8, 9, 15, 16 from depth level of 140.0 m; no. 11 from a depth level of 141.8 m): (8) *Cyclopileatispora gibberulis* (Luber) Oshurkova, comb. nov.; (9) *Raistrickia* sp.; (10) *Florinites pumicosus* (Ibrahim) Schopf, Wilson et Bental; (11) *Florinites luberae* Samoilovich; (12) *Florinites visendus* (Ibrahim) Schopf, Wilson et Bental; (13) *Plicatipollenites indicus* Lele; (14) *Luberisaccites subrotatta* (Luber) Dibner; (15) *Limitisporites moersensis* (Grebe) Klaus; (16, 17) yellow-green botryococcalean algae (magnification 500 for all specimens).

Plate II

Table 2. Correlation of the Lapcha and Botuoba formations (eastern margin of the Tunguska syneclise) with Upper Carboniferous–Lower Permian deposits of the Kuznetsk coal basin on the basis of palynological data

Tunguska syneclise (this work)				Kuznetsk coal basin (Dryagina in *Verkhnii* ..., 1988)			
Formation	Subformation	Palynoflora	Age	Palynological zone	Horizon	Stage	Series
Botuoba	Upper Botuoba	IV	P_1^2	*Cordaitina rotata – Crucisaccites ornata – Azonaletes irregulariplicatus* (CCA)	Ishanovsk	Artinskian	P_1
	Lower Botuoba	III	P_1^1	*Cordaitina uralensis – Plicatipollenites diffusus* (CP)	Intermediate	Asselian–Sakmarian	
		II	P_1^1				
Lapcha		I	C_3	Remysporites *psilopterus – R. Mirabilis* (RR)	Alykaevsk	Kasimovian–Gzhelian	C_3

Taenisporites labdacus Klaus (0–1.0%). The general proportion between spores (up to 40%) and the pollen (up to 60%) is like that in the upper part of the lower Botuoba Subformation. The described spectrum of palynoflora definitely suggests that the upper Botuoba Subformation is of the Early Permian age (Plate II, nos. 15–21).

DISCUSSION

The succession of studied palynofloras and taxonomic changes across the Carboniferous–Permian boundary correspond in general to these characteristics of deposits in the Kuznetsk coal basin (Table 2), whose section is accepted as the standard for the Angarida region of paleoflora (*Verkhnii* ..., 1988; *Kuzbass* ..., 1996).

For instance, palynoflora from the Lapcha Formation shows the predominance of spores *Psilohymena psiloptera* (Lub.) Hart et Harr, the constant presence of monosaccate pollen of form genera *Cordaitina, Florinites, Plicatipollenites*, and the appearance of sporadic grains of bisaccate, costate, and colpate pollen. The same features are typical of palynofora from the Alykaevsk Horizon of the Kuznetsk basin, where it spans the time range of the *Remysporites psilopterus* Zone after Dibner (1976, 1977, 1978).

The dominant component of palynoflora from the lower Botuoba Subformation is monosaccate pollen–rich in *Cordaitina* forms (up to 42%) such as *C. rugulifer* (Lub.) Samoil., *C. uralensis* (Lub.) Samoil., *C. rotata* (Lub.) Medv., and *C. abutiloides* (Andr.) Dibn. They associate with the pollen of *Florinites, Plicatipollenites, Potonieisporites*, and *Bascanisporites* (up to 35% in sum). Colpate pollen of *Entylissa* constitutes up to 1.0% of this assemblage, and *Disaccites* pollen, including its costate forms, is as abundant as 3.0%. All these features show that this palynoflora is similar to the pollen assemblage from the intermediate horizon of the Kuznetsk basin and corresponds to the *Cordaitina uralensis–Plicatipollenites diffusus* (CP) Zone of the Lower Permian in palynological zonation by L.L. Dryagina (*Verkhnii* ..., 1988).

Palynoflora from the upper Botuoba Subformation is comparable with the previous one. *Cordaitina* pollen is still dominant at this level (up to 40%) and includes, in addition to aforementioned species, *C. punctata* (Lub.) Hart. The permanent components of palynoflora are *Crucisaccites ornata* (Samoil.) Dibn., *Plicatipollenites gondwanensis* Lele, *Bascanisporites verus* (Sadk.) Dibn., *B. varians* (Sadk.) Dibn., *B. stipticus* (Lub.) Oshurk., and *Florinites luberae* Samoil. In sum, monosaccate pollen represents up to 55% of the assemblage. Bisaccate pollen, its costate forms included, was also encountered at this level, where *Alisporites* forms appear for the first time. *Azonaletes* pollen grains without a dehiscence cleft and of doubtful taxonomic affinity are rather abundant here (up to 8.0%). All these

Plate II. Basic palynomorphs of palynofloras III and IV

(1–14) Palynoflora III from the upper beds of the lower Botuoba Subformation (P_1^1) (Well 530/3, no. 1 from a depth level of 33.2 m and nos. 1, 5, 10, 14 from a depth level of 21.8 m; Well 569/3, nos. 2, 3 from a depth level of 4.0 m and no. 7 from a depth level of 78.0 m; Well 140/895, no. 4 from a depth level of 77.0 m and nos. 8, 9, 13 from a depth level of 78.0 m; Well 76/395, nos. 6, 11 from a depth level of 97.0 m; Well 120/570, no. 12 from a depth level of 210.4 m): (1) *Leiotriletes pyramidalis* (Luber) Allen; (2) *Lophotriletes resistens* (Luber) Medvedeva; (3) *Acanthotriletes rigidispinosus* (Luber) Ischenko; (4) *Lycospora* sp.; (5) *Samoilovitchisaccites* sp.; (6) *Cordaitina convallata* (Luber) Samoilovich; (7) *Bascanisporites verus* (Sadkova) Dibner; (8) *Bascanisporites varians* (Sadkova) Dibner; (9) *Crucisaccites ornatus* (Samoilovich) Dibner; (10) *Marsupipollenites retroflexus* Luber; (11) *Limitisporites monstruosus* (Luber) Hart; (12) *Illinites* sp.; (13) acritarchs *Tessellaesphaera tessellata* Foster; (14) acritarchs *Tasmanites* sp. (15–21) Palynoflora IV from the upper Botuoba Subformation (P_1^1) (Well 120/570, nos. 15, 16, 19 from a depth level of 163.4 m; Well 230/910, nos. 17, 20 from a depth level of 124.0 m, no. 18 from a depth level of 129.2 m, no. 21 from a depth level of 133.9 m): (15) *Plicatipollenites gondwanensis* (Balme et Hennelly) Lele; (16) *Cordaitina rugulifer* (Luber) Samoilovich; (17) *Illinites* sp.; (18) *Vittatina costabilis* Wilson; (19) *Limitisporites moersensis* (Grebe) Klaus; (20) *Cordaitina abutiloides* (Andreeva) Dibner; (21) *Lueckisporites* sp. (magnification 500 for all specimens).

characteristics indicate that palynoflora from the upper Botuoba Subformation is similar to that from the Ishanovsk Horizon of the Kuznetsk basin and attributable to the *Cordaitina rotata–Crucisaccites ornata–Azonaletes irregulariplicatus* (CCA) Zone of the Lower Permian in zonation by L.L. Dryagina (*Verkhnii ...*, 1988).

NOMENCLATURE PROBLEMS OF PALYNOLOGICAL ZONATION

Considering that palynological zonation is highly important for stratigraphy, I would like to mention that experts in palynology should be careful with taxonomic terms used in the nomenclature of distinguished zones. In the guidebook to biozonation, T.N. Koren' notes: "Criteria, which are most important to recognize biozone, are the diagnostic assemblage characterizing it and the lower boundary delimiting the subdivision ...Only the combination of these two criteria (assemblage and boundary) delineates the stratigraphic range of a zone and the precision of subsequent correlation" (*Biozonal'nyi ...*, 1995, p. 5).

Palynological zones, established in the Kuznetsk basin and accepted as standard in the Upper Paleozoic zonation for the Angarida phytogeographic province, are characterized well with respect to their palynological assemblages; however, the second criterion of the lower zonal boundary used to be ignored in the admitted nomenclature. Names of palynological zones usually combine taxonomic terms indicating just the miospores predominant in the pollen assemblage of a zone. It seems to be more correct to term zones using first the name of the index species most abundant within the zone, and second, the name of taxon, whose first occurrence level corresponds to the lower boundary of the zone. In this case, the zonal nomenclature would satisfy both criteria defining the range of the biozone. For example, the *Remysporites psilopterus* Zone of the Alykaevsk Horizon (Dibner, 1978) should be termed as the *Psilohymena psiloptera–Marsupipollenites tunguskensis* (PT)[1] Zone. By way of comparison, the *Cordaitina uralensis–Plicatipollenites diffusus* (CP) Zone of the intermediate horizon (Dryagina in *Verkhnii ...*, 1988) should be termed as *Cordaitina uralensis–Cyclogranisporites asperatus* (UA) Zone, and the *Cordaitina rotata–Crucisaccites ornata–Azonaletes irregulariplicatus* (CCA) Zone as *Cordaitina rotata–Crucisaccites ornata–Tuberculispora tumulosa* (ROT) Zone. I hope that the above considerations will trigger a revision of nomenclature of Carboniferous and Permian palynozones established in the Kuznetsk basin and used as a standard for correlation of Upper Paleozoic continental deposits all over the Angarida.

[1] Index of palynological zone must correspond to abbreviated species, but not generic names of taxa as it was done by L.L. Dryagina (*Verkhnii ...*, 1988).

SYSTEMATICS

Having no possibility to describe here all the studied palynomorphs, I present below only the diagnostic features of five form genera, which include species widespread in Upper Paleozoic deposits of the Angarida phytogeographic province. The presented new species combinations of genera *Cyclobaculisporites, Horriditriletes, Cyclopileatispora*, and *Pileatisporites* are grouped in accordance with morphology and ornamentation type corresponding to the genus diagnosis. Species of genus *Psilohymena* are distinct from others by having a wide equatorial zone and by exine delamination near the equator of miospores.

Anteturma *Proximegerminantes* Potonie, 1970 (= *Sporites* H. Potonie, 1983)

Turma *Triletes* (Reinsch, 1881) Potonie et Kremp, 1954

Supra-subturma *Acavatitriletes* Dettmann, 1963

Subturma *Azonotriletes* (Luber, 1935) Dettmann, 1963

Infraturma *Apiculati* (Bennie et Kidston, 1886) Potonie, 1956

Genus *Cyclobaculisporites* Bharadwaj, 1955

Spackmanites—Habib, 1966, p. 637.

Type species: *C. grandiverrucosus* (Kos.) Bharad.—Bharadwaj, 1955.

Diagnosis. Radial-trilete spores of the rounded to triangular-rounded morphology have simple straight-rayed dehiscence cleft with rays extending over 2/3 of the spore radius. The area is indistinct, and exine is medium thick. Under deep focus, the baculate sculpture looks like a reticular ornamentation. Ektexine bacula cover the entire spore surface richly and homogeneously. Bacula are of various height (short or tall) in different specimens but identically tall, though varying in width, in each one. Basal diameter of a given baculum is either equal to, or slightly lesser than the top diameter. Most thick bacula can be as wide as a double height parameter. Upper baculum ends are either flat or rounded. Embossed sculptural elements are responsible for the uneven contours of spores.

Size range is 40–100 μm.

Species composition:

(1) *Cyclobaculisporites (Verrucosisporites) facierugosus* (Loose) Oshurkova, comb. nov.—Loose, 1934, p. 155, Pl. 7, Fig. 26; Germany, Ruhr basin, Carboniferous, Westphalian B (42–56 μm);

(2) *C. (Verrucoso–sporites) tuberosus* (Knox) Oshurkova, comb. nov.—Knox, 1950, p. 319, Fig. 236; Scotland, Carboniferous (100 μm);

(3) *C. (Lophotriletes) atratus* (Naum.) Oshurkova, comb. nov.—Naumova, 1953, p. 123, Pl. 18, Fig. 17; Russian platform, Upper Devonian, Famennian (40–45 μm);

(4) *C. (Angaropteritriletes) trichacanthus pars* (Lub.) Lub.—Luber, 1955, p. 52, Pl. 2, Fig. 40 (but not Figs. 39, 41 and not *Azonotriletes trichacanthus* Lub.—Luber and Walz, 1941, p. 55, Pl. 12, Fig. 188); Kazakh-

stan, Ekibastuz basin, Middle Carboniferous (50 μm) [= *Azonotriletes trichacanthus* Lub.—Andreeva, 1956, p. 258, Pl. 56, Fig. 79 or = *Lophotriletes trichacanthus* (Lub.) Naum.—Medvedeva, 1960, p. 21, Pl. II, Fig. 5a but not Figs. 5, 5b];

(5) *C. (Punctatisporites) grandiverrucosus* (Kos.) Bharad.—Bharadwaj, 1955, p. 123, Pl. 1, Fig. 1; Germany, Saar basin, Carboniferous, Stephanian C (70–90 μm);

(6) *C. (Verrucosisporites) sinensis* (Imgr.) Bharad.— Imgrund, 1960, p. 161, Pl. 14, Figs. 47, 48; China, Lower Permian (70–90 μm);

(7) *C. (Spacmanites) ellipticus* (Hab.) Oshurkova, comb. nov.—Habib, 1966, p. 638, Pl. 105, Figs. 17, 18; USA, Pennsylvania, Carboniferous, Westphalian D (50–67 μm);

(8) *C. muticus* Drjag.—*Upper Paleozoic of Angarida*, 1988, p. 139, Pl. 69, Fig. 8; Russia, Kuznetsk basin, Upper Permian (70–80 μm).

Comparison. From other genera of the infraturma *Apiculati*, which also include rounded spores, the genus in question differs by its peculiar ornamentation. Sculptural bacula of the genus are stylar, in contrast to analogous elements of other genera: granular in *Cyclogranisporites* Pot. et Kr., tubercular in *Tuberculispora* Oshurk., verrucose in *Verrucosisporites* (Ibr.) Smith et Butt., pilate in *Cyclopileatispora* Oshurk., apiculate in *Apiculatisporis* (Pot. et Kr.) Oshurk., spinose in *Spinosisporites* (Lub.) Oshurk., and piliform in *Capillatisporites* Lub.

Remarks. I include genus *Spacmanites* Hab. into the synonymy, because its morphological features, namely, rounded contours and baculate sculpture entirely covering miospores, fit well to the diagnosis of the described taxon. However, bacula of the former are more highly and densely arranged than could be considered as a feature of the species level.

Distribution: Famennian Stage of the Devonian in Russia; Carboniferous of Germany, Kazakhstan, Russia, United States, and Scotland; Permian of China and Russia.

Genus *Horriditriletes* Bharadwaj and Salujha, 1964, emend. Oshurkova

Type species: *H. curvibaculosus* Bharad. et Sal., Bharadwaj and Salujha, 1964.

Diagnosis. Radial-trilete spores of a triangular to triangular-rounded shape; sides are straight, slightly concave or convex, and the angles between them are widely rounded. The dehiscence cleft is simple, with straight rays extending over 1/2 or almost throughout the spore radius. Exine is thick. Cylindrical bacula, the main elements of ornamentation, are equally wide all along their length. Upper ends of bacula are rounded or flat. The height and occurrence frequency of bacula varies in each species. Sculptural elements cover the entire surface of miospores. Embossed elements make their contours uneven. Size range is 20–80 μm.

Species composition:

(1) *Horriditriletes (Azonotriletes) trichacanthus* (Lub.) Oshurkova, comb. nov.—Luber and Walz, 1941, p. 55, Pl. 12, Fig. 188; Russia, Kuznetsk basin, Middle–Upper Carboniferous (60–80 μm);

(2) *H. (Acanthotriletes) ramosus* (Balme et Henn.) Bharad. et Sal.—Balme and Hennelly, 1956, p. 249, Pl. 3, Figs. 39–41; Australia, Permian (26–45 μm);

(3) *H. curvibaculosus* Bharad. et Sal.—Bharadwaj and Salujha, 1964, p. 193, Pl. 2, Figs. 34–39, India, Upper Permian (26–40 μm);

(4) *H. brevis* Bharad et Sal.—Bharadwaj and Salujha, 1964, p. 194, Pl. 2, Figs. 41, 42; India, Upper Permian (20–24 μm);

(5) *H. (Acanthotriletes) hirtellus* (Andr. et Drjag.) Oshurkova, comb. nov.—*Upper Paleozoic of Angarida*, 1988, p. 144, Pl. 69, Fig. 21; Russia, Kuznetsk basin, Upper Permian (35–40 μm).

Comparison. In its sculpture, the genus is similar to *Cyclobaculisporites* Bharad., but differs from the latter by the triangular contour of miospores. As compared to other genera of the infraturma *Apiculati*, which are triangular in shape, the genus shows the peculiar baculate ornamentation.

Remarks. Palynomorphs included in the genus and extending its initial diagnosis have slightly convex sides and cleft rays almost equal to the spore radius; their sculpture consists of low, closely spaced bacula. These features characteristic of individual species are insufficient to class them with an independent genus.

Distribution: Middle and Upper Carboniferous of Kazakhstan and Russia; Permian of Australia, India, and Russia.

Genus *Cyclopileatispora* Oshurkova, gen. nov.

Genus name: from Greek *cyclos*, circle, and from Latin *pileatus*, having a cap.

Type species: *C. gibberula* (Lub.) Oshurk.—Luber and Walz, 1941.

Diagnosis. Radial-trilete spores of a rounded shape have simple dehiscence clefts with straight rays as long as 2/3 of the spore radius. The area is indistinct. Exine is medium thick, with pilate ornamentation. Cylindrical elements of the sculpture (pila) are sometimes very low, have swollen round ends, and resemble club (pestle or fungus) in shape. They are usually equal in dimensions and cover the spore surface thickly and evenly. Their clavate shape is seen along the spore contour only, because in normal projections they look like rather correct circles. The spore contour is uneven. Size range is 36–85 μm.

Species composition:

(1) *Cyclopileatispora (Azonotriletes) gibberula* (Lub.) Oshurkova, comb. nov.—Luber and Walz, 1941, p. 55, Pl.12, Fig. 189; Russia, Tunguska basin, Upper–Middle Carboniferous (81 μm);

(2) *C. (Punctatisporites) corona* (Pepp.) Oshurkova, comb. nov.—Peppers, 1964, p. 33, Pl. 4, Figs. 16, 17; United States, Illinois, Upper Carboniferous, Pennsylvanian, (68.1–85.6 μm);

(3) *C. (Punctatisporites) patula* (Pepp.) Oshurkova, comb. nov.—Peppers, 1964, p. 32, Pl. 4, Figs. 11, 12;

United States, Illinois, Upper Carboniferous, Pennsylvanian (36.0–49.6 μm).

Comparison. The peculiar shape of sculptural elements distinguishes this genus from others belonging to the infraturma *Apiculati*, which also includes rounded miospores. Ornamentation is similar to that of *Pileatisporites* Pepp., but spores of the latter are less distinctly rounded.

Remarks. The new form genus is introduced, because the species of rounded pilate miospores were described, but a genus with this features has never been mentioned in publications.

Distribution: Middle–Upper Carboniferous of Russia and United States; Permian of Russia.

Genus *Pileatisporites* Peppers, 1970.

Type species: *P. aequus* Pepp.— Peppers, 1970.

Diagnosis. Radial-trilete spores of a triangular shape have straight or almost straight sides and widely rounded angles. The dehiscence cleft is simple, with straight rays as long as 2/3 of the spore radius. The exine is moderately thick. Pilate ornamentation is formed by club-like elements equal in size and thickly covering the whole surface of miospores. Contours of the latter are rugged because of the embossed sculptural elements. Size range is 40–69 μm.

Species composition:

(1) *Pileatisporites aequus* Pepp.— Peppers, 1970, p. 101, Pl. 6, Figs. 21–23; United States, Illinois, Upper Carboniferous, Pennsylvanian (61.4–69.2 μm);

(2) *P. (Neoraistrickia) clavulatus* (Drjag.) Oshurkova, comb. nov.—*Upper Paleozoic of Angarida*, 1988, p. 143; Russia, Kuznetsk basin, Upper Permian (40–50 μm).

Comparison. As compared to genera of infraturma *Apiculati*, which include triangular miospores, the genus under consideration is distinct because of its peculiar pilate ornamentation.

Distribution: Upper Carboniferous of the United States; Upper Permian of Russia.

Supra-subturma *Cavatitriletes* Oshurkova supra-subturma nova

The group consists of form genera of dispersed miospores with a trilete dehiscence cleft, whose lacinate exine exhibits cavity (cavitary trilete spores).

Subturma *Zonocavatitriletes* Oshurkova et Pashkevich, 1990

Infraturma *Zonocavati* Oshurkova et Pashkevich, 1990

Genus *Psilohymena* Hart et Harrison, 1973

Type species: *P. mirabilis* (Lub.) Hart et Harr— Luber and Walz, 1938.

Diagnosis. Cavitary trilete spores have a wide equatorial zone and are oval to triangular-oval in shape. The dehiscence cleft is simple, with straight rays as long as the central body radius; the cleft is sometimes indistinct. Near the equator of distal and proximal hemispheres, the exine is lacinate, remaining intact close to the contact area of the proximal side and near the distal pole. The thin ektexine forms a wide equatorial zone and is frequently deformed into radial plicae at the boundary with central body; intexine of central rounded body is moderately thick. Ornamentation is lacking. Infragranular structure of the central body is locally seen. Spore contour is even. Size range is 50–105 μm.

Species composition:

(1) *Psilohymena (Zonotriletes) mirabilis* (Lub.) Hart et Harr.—Luber and Walz, 1938, p. 28, Pl.6, Fig. 75; Kazakhstan, Karaganda basin, Lower Carboniferous (50 μm);

(2) *P. (Zonotriletes) psiloptera* Hart et Harr.—Luber and Walz, 1941, p. 58, Pl. 12, Fig. 198; Russia, Kuznetsk basin, Upper Carboniferous (80–90 μm);

(3) *P. (Zonotriletes) millegrana* (Lub.) Oshurkova, comb. nov.—Luber and Walz, 1941, p. 43, Pl. 9, Fig. 139 (= *Filicizonotriletes millegranus* forma *triangulata* Lub.); Kazakhstan, Karaganda basin, Lower Carboniferous (70–80 μm);

(4) *P. (Zonotriletes) ruminata* (Lub.) Oshurkova, comb. nov.— Luber, 1955, p. 67, Pl. 7, Figs. 145, 146 (= *Zonotriletes mirabilis* forma *ruminata* Lub.); Kazakhstan, Karaganda basin, Lower Carboniferous (50 μm).

Comparison. The genus differs from *Auroraspora* (H., St. et M.) since its miospores have wider and thinner equatorial zone. In addition, miospores as a whole and their inner bodies are more rounded and have no lips around the dehiscence cleft.

Remarks. Hart and Harrison (1973) distinguished the genus in palynoflora of the Kat Formation of the Tunguska basin; unfortunately their diagnosis had only a limited print run. Miospores of the genus are widespread in Carboniferous and Permian deposits of the Angarida phytogeographic region and in the Kazakhstan province of the Euramerian region. Sometimes, they have been mistaken for either *Cordaitina* pollen (*C. psiloptera*, Medvedeva, 1960, p. 38) or *Remysporites* spores (*R. psiloptera*, Luber, 1966, p. 31).

Distribution: Carboniferous of Kazakhstan and Russia.

CONCLUSION

The studied rock samples and analysis of composition and abundance rates of pollen and spore species macerated from the rocks are used to refine the geological ages of the Lapcha (C_3) and Botuoba (P_1) formations in the eastern margin of the Tunguska syneclise. The analyzed palynofloras are correlated with coeval palynological assemblages from deposits of the Kuznetsk coal basin. It should be noted that the Carboniferous–Permian boundary is placed at the level of transition from the first to second of the distinguished palynofloras in full accordance with the accepted criteria of using palynological data in stratigraphy (*Prakticheskaya* ...,

1990). The revealed possibility to correlate the studied palynofloras with Upper Carboniferous and Lower Permian palynological zones of the Kuznetsk basin convincingly demonstrates that these zones are important for the interregional correlation of Upper Paleozoic deposits in the Angarida phytogeographic region.

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Cosmopolitan Foraminiferal Assemblages from Jurassic Deposits of Central Turkey

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Abstract—Foraminiferal assemblages from the Jurassic Alacaatli section of Central Turkey are studied. The section of the reference type for the Ankara region represents a transgressive rock succession spanning the time range from the Upper Jurassic to the Valanginian. The Jurassic carbonate–siliceous rocks of the section are characterized by diverse foraminiferal assemblages that include both benthic and planktonic species. The occurrence of characteristic benthic foraminiferal genera and species, together with the zonal planktonic species *Globuligerina oxfordiana* (Grigelis) allows these deposits to be correlated within the Oxfordian in general and to distinguish separate substages and foraminiferal zones. The peculiar feature of the considered assemblages is their composition represented by cosmopolitan forms without any endemic Tethyan species. Paleoecological peculiarities of these biocoenoses indicate that foraminiferal communities lived in deep open-marine environments. This is also confirmed by the composition of radiolarian assemblages coexisting with foraminifers.

Key words: *foraminifers, radiolarians, Jurassic deposits, Eastern Mediterranean, Turkey*

Jurassic deposits are not widespread in the eastern Mediterranean. Their natural outcrops are distributed in a mosaic manner within northeastern Egypt, the Sinai Peninsula, Israel, western Syria, central and western Turkey, Jordan, Kuwait, Iran, and Iraq. In most of the sections, the deposits are represented by carbonate, terrigenous–carbonate, and, less commonly, by siliceous and siliceous–carbonate rocks. In the Alacaatly section located near Ankara, the latter rocks contain diverse foraminiferal assemblages (Fig. 1), which are discussed in the paper. Samples from the section were collected by N.Yu. Bragin, who thoroughly studied and described them during his field work in Turkey in 1994.

The study of Jurassic foraminifers in the Eastern Mediterranean (Syria, Israel, Turkey) showed that the taxonomic composition of foraminiferal assemblages depends on bathymetric zones of the past basin, temperature, and lithology, i.e., on the main abiotic factors controlling their distribution (Kuznetsova, 1994, 1996, 1998; Kuznetsova and Dobrova, 1995, 1997; Kuznetsova *et al.*, 1996; Voznesenskii and Kuznetsova, 1999). It was established that foraminifers dwelling in the Tethyan seas during the Jurassic can be subdivided, according to their biocoenoses, into three types: endemic, cosmopolitan, and transitional. The latter are generally much less diverse than assemblages of the first two types. It was also established that endemic foraminiferal assemblages composed of the agglutinated, so-called "larger foraminifers" (complex lituolids, pfenderinids, and cyclamminids) are dwellers of the shallowest zones of the basin, i.e., of the littoral and upper sublittoral zones. As for cosmopolitan assemblages represented by high-tolerant secretory and agglutinated forms, they populated the relatively deep settings and, correspondingly, were adapted to colder water environments of the lower sublittoral and epibathyal zones. The considered Jurassic foraminiferal assemblages from the Alacaatli site undoubtedly were of this type and lived in environments of the open marine basin with normal salinity, as it is evident from the occurrence of planktonic species *Globuligerina oxfordiana* (Grigelis) (Plate, nos. 14–18). It should be noted that this typical index species of the middle Oxfordian is recorded in the studied section in extreme abundance that was never observed before. Shells of this species are perfectly preserved and provide an opportunity to study both their general structure and peculiar details of the surface and wall sculpture (Plate, nos. 14, 15). The assemblage includes, along with planktonic foraminifers, benthic species, most characteristic of which are *Paleogaudryina varsoviensis* (Bielecka and Pozaryski), *Eoguttulina simplex* (Terquem), *Ophtalmidium strumosum* (Gumbel), *Cornuspira intervacare* Azbel, *Trocholina transversarii* Paalzow, *Paulina furssenkoi* Grigelis, *Epistomina nemunensis* Grigelis, *E. multialveolata* Grigelis, and *Ceratolamarckina adiposa* Azbel (Plate, nos. 1–13). All the listed species are widespread in the middle (some of them also in the lower) Oxfordian of western Europe, the Russian platform, the Baltic region, Portugal, Madagascar, and the Atlantic shelf of Canada. In the foraminiferal zonal scale correlated with the ammonite zonation, this assemblage corresponds to the *Ophtalmidium strumosum–Lenticulina brestica* Zone (Gri-

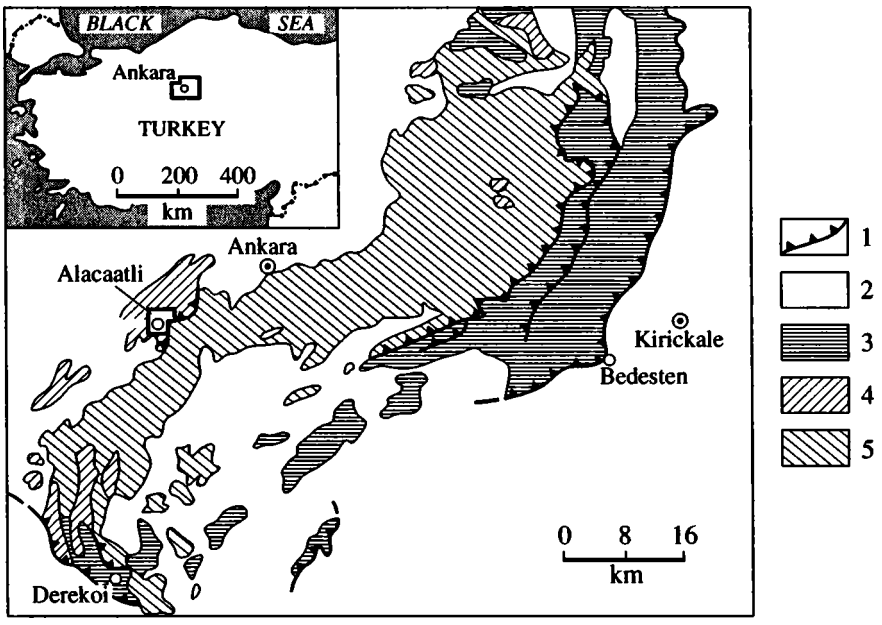


Fig. 1. Geological scheme of the Ankara area: (1) thrust faults; (2) upper Campanian, Maastrichtian, and Cenozoic deposits; (3) Late Cretaceous ophiolitic melange; (4) Jurassic and Lower Cretaceous carbonate–siliceous deposits; (5) Early Mesozoic melange.

gyalis, 1985). This assemblage of benthic and planktonic foraminifers lived in the open sea basin with the normal salinity, which is also confirmed by the radiolarian assemblage occurring in beds underlying and overlying the foraminifera-bearing deposits. A detailed description of the section located near the Alacaatli Village (Fig. 2) has been provided by Bragin and Tekin (1998), therefore we discuss here only the Upper Jurassic part of the section from the deposits directly underlying beds with foraminifers to the member overlying there beds.

East of the Alacaatli Village along the right-hand side of the Derigendere Ravine, the following succession of beds is exposed (Fig. 2):

Bed 1: cherts, limy, rose to yellowish and greenish gray, platy, yielding radiolarians (Sample 1) *Acanthocircus trizonalis trizonalis* (Rust), *Emiluvia sedecimporata* (Rust), *E. salensis* Pessagno, *Mirifusus dianae* (Karrer), *Podocapsa amphitreptera* Foreman, *Ristola altissima* (Rust), *Tetratrabs* sp. cf. *T. bulbosa* Baumgartner, *Triacoma* sp. cf. *T. blakei* (Pessagno), *Tritrabs* sp. cf. *T. casmaliaensis* (Pessagno), *T. ewingi* (Pessagno), and *T.* sp. cf. *T. exotica* (Pessagno), which indicate the early Oxfordian–early Tithonian age of the host deposits (Baumgartner *et al.*, 1995). The apparent thickness is 3 m.

Bed 2: limestone, dark-gray, massive. The thickness is 1 m.

Bed 3: limestones, red and rose gray, cloddy, with obscure bedding, sandy, lacking graded textures, bearing ammonite and belemnite remains. The bed encloses foraminifers (Sample 4) *Cornuspira intervacare* Azbel,

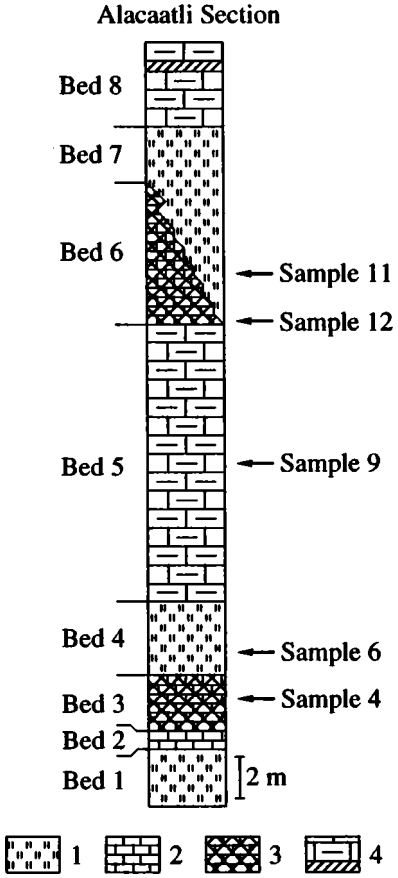
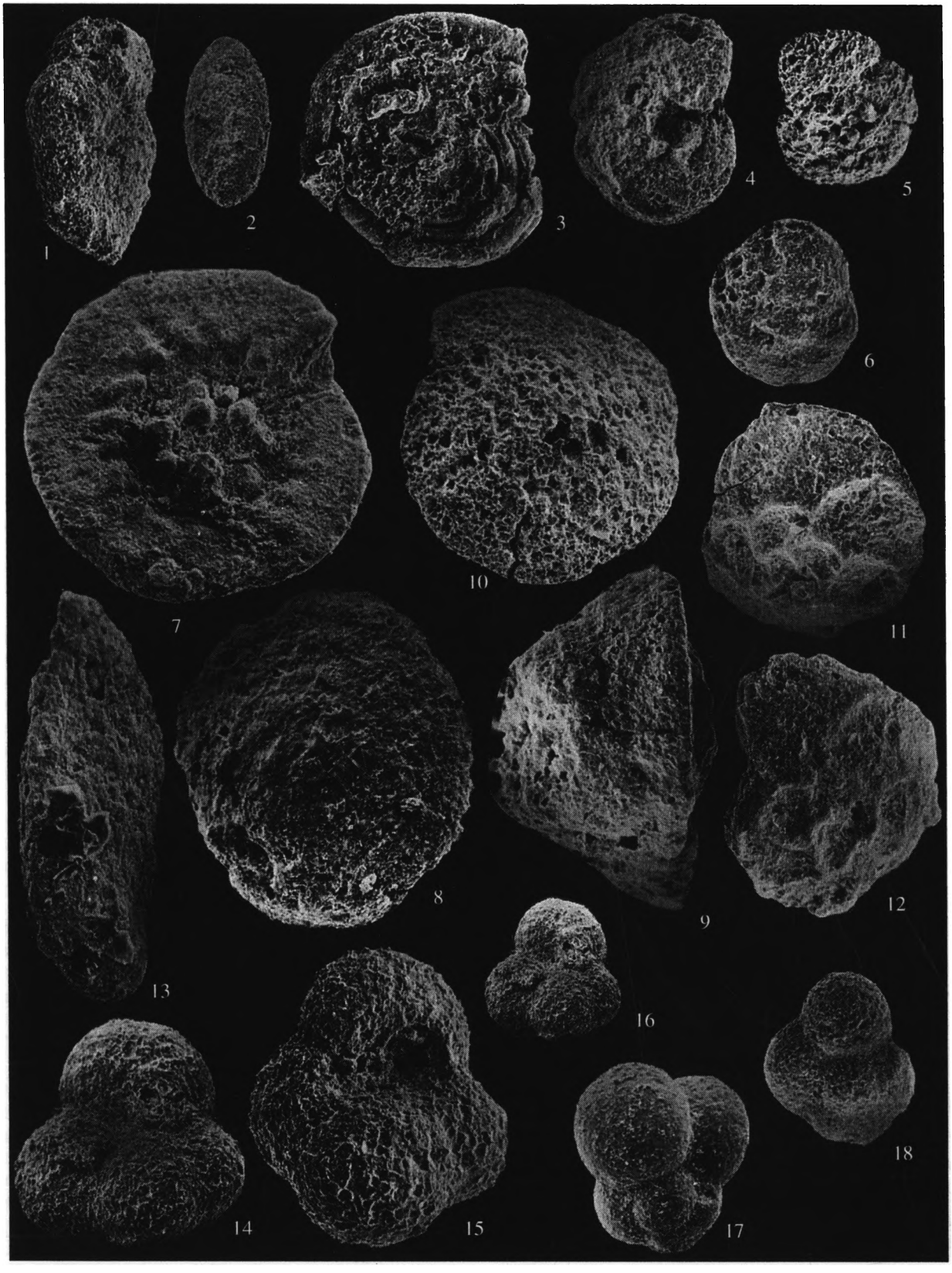


Fig. 2. Structure of the Jurassic–Cretaceous section near the Alacaatli Village and location of beds with foraminifers: (1) chert; (2) micritic limestone; (3) limestone of the "ammonitico rosso" facies; (4) clayey limestone and marl.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Paleogaudryina varsoviensis (Bielecka and Pozaryski), *Eoguttulina simplex* (Terquem), *Ophtalmidium strumosum* (Gumbel), *Ceratolamarckina adiposa* Azbel, *Paulina furssenkoi* Grigelis, *Epistomina nemunensis* Grigelis, *E. multialveolata* Grigelis, *Trocholina transversarii* Paalzow, and *Globuligerina oxfordiana* (Grigelis). The assemblage is characteristic of the *Ophtalmidium strumosum Zone* of the middle Oxfordian (*Biostratigrafiya...*, 1982). The thickness is 3 m.

Bed 4: cherts, variegated, banded red-yellow-green, slightly calcareous, platy, with thin (to 0.5 m) intercalations of gray massive limestone. The bed contains all the radiolarian species that are recorded in Bed 1 in association with (Sample 6) *Acalniotyle diaphorogona* Foreman, *Angulobracchia purisimaensis* (Pessagno), *Emiluvia orea* Baumgartner, *Napora bukryi* Pesagno, and *Spongocapsula perampla* (Rust). Species *Tetratrabs* sp. cf. *T. bulbosa* Baumgartner disappear in the middle of the bed (Sample 7). The thickness is 4 m.

Bed 5: limestone, clayey, gray-green, platy. The first occurrence of radiolarian species *Acaeniotyle umbilicata* (Rust) is recorded in this bed (Sample 9). The thickness is 15 m.

Bed 6: limestone, red and rose, cloddy, clayey, bearing an insignificant sand admixture, and abundant ammonite and belemnite remains. The foraminiferal assemblage of the bed (Sample 12) includes similar benthic species and *Globuligerina oxfordiana* (Grigelis) of the Oxfordian age. Limestone composes a lentil to 7 m thick and replaced, along the strike, by banded cherts with the late Oxfordian–early Tithonian radiolarian assemblage (Sample 11) similar to that from Bed 5.

Bed 7: cherts, limy, red-green, banded, platy, enclosing (Sample 13) the radiolarian assemblage similar to that from Bed 5. The last occurrence levels of *Emiluvia sedecimporata* (Rust), *E. orea* Baumgartner, *Napora bukryi* Pesagno, *Spongocapsula perampla* (Rust), *Triactoma* sp. cf. *T. blakei* (Pessagno), and *Tritrabs* sp. cf. *T. casmaliaensis* (Pessagno) are recorded within the bed. The thickness is 7 m.

The Alacaatli section is one of few Jurassic sections with coexisting radiolarians and foraminifers. Moreover, the latter are represented by very peculiar assemblages indicating the middle Oxfordian age of host rocks. The absence of foraminiferal forms redeposited or atypical of the Oxfordian deposits eliminates the possibility of a hiatus in sedimentation and reworking of fossils. It should be noted that the marine benthic community is represented here only by foraminifers; any other remains of benthic dwellers (bivalves, brachiopods, corals, ostracodes, echinoderms, and others) are absent. This phenomenon needs some explaining, which should be evidently found among the abovementioned abiotic factors, first of all, in bathymetric and temperature characteristics of this part of the Late Jurassic sea basin.

In contrast to many Jurassic successions investigated in the central and marginal Tethyan regions, e.g. in Syria, Israel, Egypt, Saudi Arabia, Kuwait, Yemen, Iraq, Iran, Jordan, Morocco, Crimea, Lesser Caucasus, Central Asia (Kuznetsova and Gorbachik, 1985; Kuznetsova *et al.*, 1996; Alsharhan and Nairn, 1997), only the Alacaatli section of Turkey includes red limestone beds of the ammonitico rosso facies, which enclose foraminiferal assemblages. This sufficiently widespread deep-water facies is mainly typical of the Alpine fold belt, where its beds are recorded in the Apennine Peninsula, Balkans, and other regions. Accumulation of these sediments is usually characteristic of deep offshore zones of open sea basins with minimal terrigenous influx. Siliceous and calcareous planktonogenic sediments were accumulated very slowly, as the Oxfordian deposits in the Alacaatli section are less than 40 m thick. It should be remembered that the thickness of Oxfordian rocks in Syria amounts to 290–300 m and that in Crimea varies from 650 m in the Sudak synclinorium to 1800–1900 m in eastern Crimean synclinoria and in western Crimea.

It is assumed that the ammonotico rosso facies accumulated on small local underwater rises in the deep (epibathyal–pseudoabyssal) part of the Jurassic basin, which was inhabited only by pelagic organisms including pelagic bivalves recorded in Oxfordian sections of Italy (Sartorio and Venturini, 1988) and Carpathian Mts. (Patrunov, 1988).

It is interesting that only foraminifers could adapt and develop on the sea bottom under such conditions that none of many Jurassic benthic faunas could survive. Foraminiferal assemblages from the considered section are identical, at the generic and species levels, to those from the shallow epicontinental seas of the Russian platform, relatively deep shelf basins of Madagascar, the Atlantic shelf of Canada, and from deeper deposits of Portugal, the Baltic region, Poland, Germany, England, and other localities. This suggests a

Jurassic foraminifers from the Alacaatli section (all pictured shells are from the lower–middle Oxfordian deposits of the Alacaatli section, Turkey).

(1) *Paleogaudryina varsoyensis* (Bielecka and Pozaryski), lateral view, ×250; (2) *Eoguttulina simplex* (Terquem), lateral view, ×140; (3) *Cornuspira intervacare* Azbel, lateral view, ×250; (4) *Paulina furssenkoi* Grigelis, ventral view, ×140; (5) *Epistomina* cf. *nemunensis* Grigelis, ventral view, ×140; (6) *Paulina furssenkoi* Grigelis, dorsal view, ×250; (7–9) *Trocholina transversarii* Paalzow: (7) ventral view, (8) dorsal view, (9) lateral view, ×140; (10) *Epistomina multialveolata* Grigelis, dorsal view, ×250; (11) *Paulina furssenkoi* Grigelis, dorsal view, ×140; (12) *Epistomina nemunensis* Grigelis, dorsal view, ×140; (13) *Ophtalmidium strumosum* (Gumbel), lateral view, ×140; (14–18) *Globuligerina oxfordiana* (Grigelis): (14) umbilical view, ×250; (15) umbilical view, ×300; (16) umbilical view, ×150; (17) lateral view, ×200; (18) spiral view, ×150.

very high tolerance for foraminifers, which migrated following the Oxfordian transgression and adapted to any environments within a wide range of depths, latitudinal–climatic zones, and lithology. This ability to survive was favorable for the large-scale distribution of foraminifers. Simultaneously, it should be noted that the life span of most species from the considered coenoses was very short, exceeding sometimes only 0.7–1 m.y. and corresponding to the part of a stage or to the foraminiferal zone. Two peculiar features of the Jurassic cosmopolitan species, namely the high tolerance degree and short life time, make this group useful for the extremely wide, even subglobal correlation of Upper Jurassic marine deposits.

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Long and Laborious Way toward the Theoretical Stratigraphy

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More than 20 years have passed since the time when S.V. Meyen had published his fundamental research work *Introduction to the Theory of Stratigraphy* (1974), which was of an irrevocable significance for the development (to be more exact, for the formation) of a poorly understood theory of stratigraphy. The latest works by Meyen and the posthumous issue of the above-mentioned book (Meyen, 1989) summed up previous studies in the field of theoretical stratigraphy. After the publication of a series of national stratigraphic codes, including the Russian Code of 1992 and second issue of the *International Stratigraphic Guide* (1994), the total number of works dealing with the problems of theoretical stratigraphy decreased substantially.

Exceptions are the fascinating papers and books by K.V. Simakov devoted to the geologic time and different stratigraphic problems and published one after another since the early and, especially, the mid-1980s. In more than half of these publications, the first theme is considered, and we may convincingly maintain that no one but Simakov analyzed this theme so comprehensively and thoroughly. His book *An Essay on History of the Concept of Real Geologic Time* (Simakov, 1996) represents a generalization of previous studies.

A new book by Simakov, *On the Way to the Theoretical Stratigraphy* (Simakov, 1997), is devoted to a wider range of theoretical problems. It is a result of the author's long-standing and partially published studies on issues concerning the historical–theoretical aspect of the problem and on the Earth's stratigraphic structure. Simakov has considerably extended the object of the book trying to find, not without success, the explanations to different stratigraphic problems in their relation to more general problems and laws of knowledge. The author believes that the fundamental stratigraphic concepts can be substantiated more reliably by this strategy. Therein is the unique nature of the book. This undoubtedly increases its value for a reader who is interested in the theory and well experienced. However, such a satiation with the cognitive material is difficult for the perception of most readers, and general definitions which are suggested, for instance, for the stratigraphic unit (pp. 69, 86), taxon (p. 84), stratigraphic subdivision (p. 69), and classification (p. 78) will hardly come into general use among stratigraphers.[1] A short list of non-stratigraphic terms given in the book assists, for sure, in the understanding of many things, but a special training and a peculiar brainwork are required to assimilate them. Having no adequate background for considering the wealth of the material presented in the book, I will discuss only stratigraphic problems that are most topical and familiar for me.

I will look at the problems in the other sequence than Simakov did in his book, applying my habitual scheme: the subject of stratigraphy, boundaries of stratigraphic units, classification and taxonomy, the time of stratigraphy, basic principles in stratigraphy, its subject-matter and definition. It is good to start from some initial points concerning peculiar features of the subject (subjects) of stratigraphic studies.

Specificity of stratigraphic information. Simakov considers the specificity of geologic information using the stratigraphic and paleontologic material as the basis. Having analyzed the works by Ch. Layelle, Ch. Darwin, G. Barrel, and I.A. Efremov, he arrived to a series of conclusions that could be formulated as follows:

(1) The supergene earth crust or its constituents ("systems of the historical record") represent *static* systems, i.e. the results of preceding dynamic processes: "of a complicated interaction of genetically different systems and processes, the influence of which is recorded in natural geologic bodies (stratigraphic units)" (p. 41).

(2) An originally incomplete geologic (stratigraphic included) record turns out to be much more incomplete in the course of time. The incompleteness of the stratigraphic record is manifested first of all in stratigraphic hiatuses and selectivity of organic remains preserved.

(3) The preservation of stratigraphic information (stratigraphic units, organic remains) depends on critical coincidences in heterogeneous processes in the lithosphere and biosphere, i.e. the information itself is of probabilistic nature (statistically probabilistic in terms of Simakov).

[1] Hereinafter, pages in the book by Simakov (1997) are quoted.

(4) The time (chronological) succession of stratigraphic units is inferable on the basis of their spatial (lower–higher) relationships.

Simakov suggests that the concept of incompleteness of the geological record should be termed as Layelle–Darwin's law, regarding it as a particular case of the fundamental "law of tempodescinence" (Meyen, 1989, p. 178).

The book is focused on the statistical geologic information turning into dynamic retrospective reconstructions on the basis of the actuality principle and gives the basic restrictions to the principle. The restrictions are caused, on the one hand, by the statistically probabilistic nature of information, and on the other hand, by a series of general peculiarities of systems. One of the peculiarities is stated as follows: "the similarity of systems does not mean that they originated in response to action of one and the same factor, under the same conditions, or as a result of both reasons" (Urmantsev, 1974, p. 90).

Subject of stratigraphy. The majority of researchers consider the stratigraphic unit as the subject of stratigraphy. This is directly or indirectly authorized in both the national stratigraphic codes and international stratigraphic guides. Simakov classifies the supergene earth crust as "the general subject" of stratigraphy and natural geologic bodies as its specific subject. It seems to be more reasonable to hold the stratigraphic unit as the subject of stratigraphy, but not the natural geologic bodies in general. And why are the latter termed as *specific* subjects of stratigraphy?

The nature of stratigraphic units and their systematic arrangement, the substantiation of which is discussed in corresponding sections of the book, are beyond of doubts at present. It is significant that the author considers chronostratigraphic (general) units in these sections as well. His other important conclusion is that stratigraphy and geology as a whole deal with the recorded development of *open* systems of different types, including the geological and biological ones. "The geologic open systems have only the energy and matter, and their evolution depends on three so-called key factors: the climatic, tectonic and hydrodynamic regimes. The biologic systems exchange their matter, energy, and information with the environment and thus belong to the type of self-organized nonlinear systems" (pp. 55–56).

This is a basic conclusion. However, I believe that it is not in agreement with Simakov's statement on autonomy of the evolution laws of genetically different systems and processes. I regard the expression "autonomy of laws" as being not clear for me, as the statement of autonomy of evolution. This is confirmed in the subsequent explanation, where the author maintains that taken for basic features distiguishing stratigraphic complexes could be "signs characterizing the changes in either the tectonic regime and structural features, or in the composition and structure of fossil assemblages, and also in paleomagnetic data. In all these cases, we will get natural geologic bodies (stratigraphic units) of different stratigraphic categories (tectono-, litho-, bio-, magneto-stratigraphic, etc.) and incompatible in their spatial–chronological (stratigraphic) range" (p. 71). This is used to conclude that the comprehensive substantiation[2] of local stratigraphic units, which is required in national stratigraphic codes, causes objections. At the same time, it is accepted in the book that "we are forced to *integrate* different methods when separating the particular regional stratigraphic complexes" (p. 71).

I should point out that the author's interpretation of the autonomy of evolution of different systems leads us exactly to the "multi-gage" stratigraphy, when all "stratigraphies" have equal rights because different methods allow us, in principle, to interpret the time–space structure related to heterogeneous paleosystems" (p. 66), but to interpret them in principle, i.e., only theoretically.

How does such a complete (judging from the previous citations) autonomy agree with the openness of stratigraphic systems experiencing mutual influence, penetrative effects, and interaction? Are those stratigraphic units, which we call basic and Simakov refers to as "classic", an integrated result of "interaction" (to different and varying degrees) between a series of dissimilar paleosystems that is recorded in diversified characteristics of a stratigraphic unit, which are of different importance in one case or another? The relative stratigraphic significance of the above charactristics is not discussed in the book, or maybe I have not noticed this issue, which, as I should mention, is of great importance (Meyen, 1974; Zhamoida, 1977).

I regard Simakov's recognition of magnito-, seismo-, and other "stratigraphies" as specific methods of the general (unified) stratigraphy as a due consideration of all characteristics different in rank. In other words, he believes that theoretically all "stratigraphies" have equal rights, but practically (and naturally) there are basic stratigraphic units representing the evolution of geo- and paleoecosystems (p. 66), and also the stratigraphic units of specific substantiation which may serve the correlation procedure. Categories of basic and special stratigraphic units in national stratigraphic codes, or basic stratigraphic units and units of different correlation methods in the British Code of 1978, represent a good example, when all characteristics depending in general of specific aspects of stratigraphic studies are taken into account (Zhamoida, 1994).

If we leave aside the indigestible nongeological terms plentiful on some pages (pp. 69, 86), then the suggested general definition of a stratigraphic unit seems to be as follows.

[2] It would be more appropriate to change the term "substantiation" for "characteristics" (Zhamoida, 1980, p. 40) in the corresponding statements of "The Stratigraphic Code" (1992, p. 35).

A stratigraphic unit is a geologic body representing a static paleosystem formed as a result of interaction between different processes in the course of certain time interval, occupying a definite position in the stratigraphic section, and showing a series of features which distinguish it from adjacent stratigraphic units situated above, below, and laterally.

Simakov correctly states that "the complete temporal coordinates of any natural phenomenon can be established only when we know its age and lifetime, which are not the same things" (p. 112). I believe the lifetime of a paleosystem corresponding to a particular stratigraphic unit was named by the author with a lame term "the frailty." In any case, all dictionaries give quite another meaning to the word "frailty." Spizharskii (1987), following V.I. Vernadsky, used more appropriate terms for this notion—duration or continuance. The words duration or continuance characterize paleosystems, although we usually deal with a geologic body (stratigraphic unit). Therefore, any definition of the latter should outline its particular position in the stratigraphic (precisely stratigraphic; see Zhamoida, 1994) section. Definitions of this kind are available in national stratigraphic codes (1997, 1992) and in international stratigraphic guides (1976, 1994). The duration or continuance of any paleosystem formation is within the time range of a stratigraphic unit that is defined in the *Stratigraphic Code* (1992, p. 21) as "the maximal interval of a geologic section between stratigraphic boundaries of the unit." "The stratigraphic range of a given formation should be evaluated in its most complete section; i.e., it should correspond to the whole time interval of rock generations enclosed into the formation" (Ibid., p. 38). This important notion concerning the range of stratigraphic unit is formulated by Krishtofovich (Krishtofovich, 1939; Zhamoida, 1980).

Simakov's attempts to define all plausible modifications of stratigraphic units and to formulate versatile criteria for their distinguishing led him to the inference that there is a possibility to use for this purpose the laws of transformation, which control all phenomena in natural systems (Urmantsev, 1974). According to the mentioned laws, there are only 15 types of system transformations defining the versatile criteria. It is likely the case, though I doubt that Table 3 (p. 73), compiled on this basis, can be useful for the practical distinguishing of stratigraphic units.

Boundaries of stratigraphic units. It was Meyen (1989) who emphasized that boundaries of stratigraphic units reflect reorganizations in corresponding paleosystems. When discussing the continuous–intermittent formation processes of natural geologic bodies (stratigrafic units), Simakov arrives at the conclusion on the nature event-related of boundaries of *any* stratigraphic (including chronostratigraphic) units irrespective of their scale. The nature of boundaries (whether they are abrupt or gradual) is determined by the genesis of stratigraphic units and by "their compliance with a particular element of a complete cycle in the evolution of a given system, whose arrangement level the units represent" (p. 71). What is said can be supported in principle. However, how objective would this interpretation be, especially the definition of an element of the complete cycle of the system evolution? The suggested definition of stratigraphic boundaries (p. 71) is suitable for any system evolving with time but not to geological systems only.

When comparing typologic and systematic approaches to the definition of chronostratigraphic boundaries (Table 6, pp. 149–150), the author shows the advantages of the latter and the transitional (ecotone) or volumetric character of these boundaries. "In such a situation, not only the paleobiologic boundary datum selected by convention (which represents only its formal nomenclature standard) should be subjected to standardization (fixation in the stratotype) but also the record of *the event* itself representing the real (not conceptual!) chronostratigraphic boundary" (p. 151). In other words, the boundary stratotype (limitotype) must include not only the global stratotype section and point (GSSP)[3] but also a set of parachronological markers which are topologically synchronous with the GSSP. In this case, the Meyen principle can be used. Simakov's proposal cited above is of practical importance for verifying and tracing the boundaries.

I would like to recall that the *Code of Stratigraphical Nomenclature for Norway* (1961) holds that the boundary between chronostratigraphic units represents only an ideal time level, whereas in practice it is proved to be a certain time interval or time-belt.

At one of the first sessions of the ISC Commission on Stratigraphic Classification and Nomenclature held in 1996, I suggested to use the "volumetric boundaries" in the geological practice for local stratigraphic units and to consider their time interval as depending on the boundary rank in the general stratigraphic scale. My report has not been the subject of much discussion and has not been published as well, but the suggestion was realized in the *Stratigraphic Code of the USSR* (1977, p. 27), where "diachronism" of stratigraphic boundaries of local stratigraphic units is considered. The volumetric boundary was accepted to be a unit of a lower rank in the general scale, rather than the rank to which the local unit is adjusted. The recommendation has not been included, however, in the second edition of the code.

Simakov's attitude to the suggestions of A.P. Karpinskii and L.L. Khalfin, who tend to distinguish "transitional layers" as *independent* units of the general stratigraphic scale, is still unclear to me. The "transitional layers" with peculiar paleontological characteristics do occur, but does it follow from this fact that there is a necessity to introduce new, "buffering" units of the

[3] In the Russian *Stratigraphic Code* (1992), the abbreviation is translated as "the point of the global boundary stratotype.

general scale *between* those universally adopted? Most geologists, justly disagree with this.

An important point is another phenomenon, termed by Simakov as morphostructural diversification, in the course of which "there acts a kind of testing of possible ways of the subsequent evolution of the ancestral biosystem (taxon) and selection of archetypes more adaptable to new conditions" (p. 47). As this takes place, we meet ephemeral taxons (the dead-end evolutionary branches) violating the gradual course of evolution. Simakov considers this phenomenon as one of the "most significant feature of biosystems, i.e., their *nonlinearity*, which differentiates them from geosystems of the deterministic type" (pp. 47, 48). The conclusion is of essence for establishing the specificity of biologic systems, paleobiosystems included. The time–space ranges corresponding to such "transitional" events can be included into the adjacent unit of the general scale. Hence, the duplication of "workable" (adopted) boundaries can be eliminated.

Unfortunately, the lateral boundaries of regional and local stratigraphic units, the importance of which was mentioned by Krishtofovich (1939), are left beyond consideration in the book. The relationship between lateral and stratigraphic boundaries of the above-mentioned units is an unresolved problem, though some considerations on the theme have been published (Krasilov, 1977; Zhamoida, 1980, 1984; and others).

Classification and taxonomy. By mentioning the lack of general definition of the stratigraphic classification, Simakov regards the last term as a "distribution of objects of a closed space as classes (orders, taxons) in accord with the measure of their identity and distinction relative to the selected basis for comparison (division)" (p. 78). He believes that classifications suggested by various researchers are not of this sort in the stratigraphy because, when distinguishing and ranging the taxons, "the principal condition of any classification, namely, the common criteria for division, is not fulfilled" (p. 76). The main "misfortune" for stratigraphers was the compilation of the taxonomic scale of stratigraphic units on the basis of hierarchy of biologic taxons that was proposed by E. Renevier in 1874.

How do classification and taxonomy stand in stratigraphy? Simakov distinguishes three trends in the elaboration of stratigraphic taxonomy: the first is based on the natural periodization of geological processes, the second follows the path of discrimination of stratigraphic units in accordance with their spatial ranges, and the third is the distinguishing of stratigraphic units on the basis of their time–space ranges corresponding to stages in the geological evolution of different-scale geosystems. The third trend represents the development of Renevier's ideas and is realized in the national stratigraphic codes. The second one is, in my view, not independent and represents a part of the third trend.

Simakov emphasizes a principal distinction between subjects of stratigraphic and biological classification. In the biological classification, these "are organisms (or their fossil remains) representing (or represented) a system, individual parts of which ("merons") cannot be self-functioning... On the contrary, merons of regional stratigraphic units (horizons) themselves represent stratigraphic units (formations); in other words they can exist independently" (pp. 94, 95).

Subjects of biologic and stratigraphic classification (organisms and stratigraphic units) differ radically. However, why do formations need to be recognized as the parts (merons) of a horizon, but not as taxons subordinate to it? If we consider formations as merons of a horizon, then horizons (stages) will be merons of series, series will be merons of systems, and only the lithosphere will be a natural subject of stratigraphy (and geology?). Actually, organisms are natural subjects of the biological classification, yet species, genera, etc. (or taxons "in accord with the measure of their identity") tangibly exist as well. The other thing is that the taxon range can be recurrently revised and studied: low-rank taxons can be redistributed among taxons of higher rank due to the progress in investigations. Yet similar redistributions (formations in horizons, stages in series, and so on) are characteristic of stratigraphy, and the time–space relationships between stratigraphic taxons represent the common basic for their subdivision, do not they? The main word here is time, by the way.

Following Krut' (1978), Simakov distinguishes two aspects of the notion "taxon": collecting and conceptual. In the biological classification, where taxon is a group of individuals possessing common features (properties), the aspects mentioned above seem to be aligned. In stratigraphy, Simakov emphasizes the conceptual aspect, which must determine the essence of the stratigraphic unit, i.e., a certain level of organization of the genetic-type systems: "solution of the stratigraphic classification cannot be obtained within the framework of stratigraphy only, classification should generalize here the results of allied scientific branches, such as paleogeograhy, lithomology, facies study, etc." (p. 95). The stratigraphic classification in his opinion, "is oriented toward the recognition of natural geological bodies characterizing some types of environments" (p. 90). As the subsequent example suggests, the beach deposits represent a type of environments and, hence, they correspond to a unit of stratigraphic classification, whereas pebbles or sand are units of the lithologic (lithomologic, in terms of V.N. Shvanov) classification (gileation) and represent "polymorphic modifications of one taxon" in the context of the first classification (p. 91). I doubt the validity of this approach to problems of stratigraphic classification.

The problem posed seems to be a point open to discussion. The stratigraphic unit is surely a geological body consisting of certain rocks formed under certain environments. Yet I wonder whether the mode and conditions of the body formation are principal or only char-

acteristic features (properties)? The main distinction of stratigraphic units from units of the lithological, formational, and facies classifications is the priority of their time or geochronologic coordinates "calculated" with the help of spatial coordinates.

Simakov suggests a system (Table 4, p. 87) of principal taxonomic categories of stratigraphic units of geo- and paleoecosystems. Units of regional and local taxonomic scales adopted in our country and hierarchically organized into the evolution cycles (period–stage–phase–rhythm, (i.e., the elementary parts of a process) are bound to different spatial subdivisions of the biosphere (e.g., the biosphere as a whole or its particular settings). The table can be regarded only as an ideal, worth striving for. In practice, however, not all formations can be "foisted" into horizons (this was taken into consideration in the *Stratigraphic Code of the USSR*), like the horizons into the regional or interregional stages. A member, which is regarded as a part of a given formation, cannot correspond to a stage, i.e. to that element of a complete evolution cycle as a formation. I am inclined to believe that the idea to regard a paleobasin as a part of a region and the region as a part of a province is incorrect. All these elements are distinguished in the book (p. 37) on the basis of different features and can easily be commersurable in their spatial ranges.

Having noted the significance of biological events for the recognition of chronostratigraphic boundaries and compilation of stratigraphic scales, Simakov considers the events in two aspects (taxonomic and biogeocoenotic) and terms them, I do not know why, as levels of biota organization. In the first case, we deal in his opinion with biological units and scales, which are proposed to be named biotemporal, but in the second one, with paleoecosystem units and scales considered as biostratigraphic.

The very complicated definition of the biotemporal scale cited above is a similarity of a linear structure that represents the morpho-structural evolution of a separate group (phylum) of organisms and is divided following its evolutionary stages (p. 133). Nonequivalence of biological taxonomic categories identically named and the dependence of biotemporal scales on selected groups of organisms lead to different age interpretations of stratigraphic units and their boundaries.

Unlike biotemporal ones, the biostratigraphic units (from the above-mentioned viewpoint) represent records of evolution of biogenic components in different-rank paleoecosystems. Thus, "Biostratigraphic scales reflecting the evolutionary stages of their biogenic component or separate fossil groups have the volumetric–hierarchic (multilevel), but not the linear structure" (p. 136).

If I correctly understood the sentences cited above, the phylozones can be attributed to biotemporal units, whereas the taxon distribution zone, acme zone, and concurrent range zone can be regarded as biostratigraphic units (*Stratigraficheskii kodeks*, 1992).

Time and Stratigraphy. Analyzing the development of concepts of the geologic time, beginning with that of N. Steno, Simakov (1996) concludes that the International Startigraphic Scale (ISS) is a "manifester" of the time. He recognizes three principal competing concepts available at present: (1) autochronological ("the geological history is a chronometer"), (2) biochronological ("the geological history has a chronometer"), and (3) typologic (the ISS is an artificial structure).

The author arrives at the conclusion that the first and second concepts cannot be adopted, as the modern ISS structure does not fit principles and criteria of distinguishing its units, which are formulated by supporters of these concepts. There is no way to agree with the assessment that the ISS is artificial. Although its units and their boundaries are natural, the scale in essence is a model biography of the Earth's hypergene crust and cannot be used "as a basis for constructing a *device* for measuring the real geologic time" (p. 109).

What is the real geological time in Simakov's opinion?

First, the time is not dynamic, perceivable as "ever flowing *continuance*" (i.e. quantititatively), but static and depicting "*qualitatively different* successive states of a single, or replacing one another material systems" (p. 119). "For the natural markers of the geologic time, we can take only the records of unique phenomena reflecting the irreversibility of paleosystem evolution" (p. 120), i.e., the qualitative transformation of material (geological) bodies.

Second, "the statistical nature of the geologic information,... excludes the key possibility to reduce the geologic time to physical (common)" (p. 119), i.e., Newtonian time.

Meyen (1989, p. 127) descriptively formulated these conclusions in the following way: "the notion 'geologic record' is proved to be surprisingly exact and not fully corresponding to the classic (Newtonian) concept of the time. One should remember that geoscientists do not have (and cannot have) any other time except for that fixed in the record."

Here we have a series of important conclusions limitating the use of some notions and methods in geology, in particular, the notion of the absolutely exact measurement and assessment of geological phenomena lifetime on the basis of the average continuance of different-rank cycles. The erroneous nature of methodological basis of the modern chronostratigraphy is also noted in the book as based on the idea of identity of the physical and geological time that was suggested by H. Hedberg (1961).

It is Simakov's opinion that the complexity of natural calibration of the geological time record lies in the fact that reliable classification of natural geologic bodies has not been actually elaborated, because taxons identically named in different higher taxonomic groups

are not equivalents. This is correct in regard to stratigraphic units distinguished by any methods: lithological, paleontological, and others.[4]

One can agree with the above-reasoning. The ISS in the present state cannot perform the functions desired originally. Simakov believes that the ISS cannot serve as a basic device for measuring the real geological time, in particular, as an instrument for measuring the paleobiologic time "because its principal aim is to fix the position of the established boundaries, but not to reconsider their status, or to revise and re-evaluate the range and taxonomic rank of the ISS units separated by these boundaries" (p. 109).

As is noted further, the creation of such a device (scale, method, calendar) is a problem in its own, and Simakov suggests principal premises for solving it using the set theory and the idea of binary relationships that is substantiated in the book.

A special background is required to understand his suggestions, but I am concerned for the call to revise the ISS. The idea is not new.

So far as the idea to revise and reorganize the ISS is concerned, it may result in a complete failure of geological cartography. To avoid this, we have the only way: all maps in use must be digitized and inputted into a single global computer system which is not available. The indefiniteness of the result is of even greater concern to me. "The practical procedure of measuring the static time is a *calculation* of simple primary measures representing the equivalence class of evolutionary stages of particular paleosystems whose rank is lowest in the general hierarchy of paleosystems of that genetic type, which is taken as basic to model the time" (p. 125). It is hard for me to even imagine an actual possibility of choosing such a simple unit of the geologic time, which would correspond to an elementary stage of the paleosystem evolution. At least, this should be approved by the majority.

Maybe it would be more correct to believe that the ISS is a result of purely regional synthesis, as it has been shown long ago by Sokolov (1971)? More constructive appears to be the idea proposed by Teslenko (1976) who used the term "integration" to explain the relation of local and regional units to general units of the ISS. The first relate to the last as quotient to common. The idea is laid to the basis of native *Stratigraphic Codes* (1977, 1992). I believe that the results of the so-called event stratigraphy and, first of all, the work by Walliser (1995) confirm to some extent the ISS validity, since the scale reflects the qualitatively different consecutive states of paleosystems. It is likely that the specificity of the real geologic time is exactly in absence of a single elementary measure. Nevertheless, my questions and doubts might be a consequence of my incompetence in such a complicated problem.

Principles and laws of stratigraphy. Thirty-three "principles" and "laws" of stratigraphy are listed in the book in the author's formulations (Table 1, pp. 14–21), and their relationship with the cognitive fundamental laws is elucidated (Table 2, p. 54). This seems to be the most complete list of postulates having a potential to be included into the theoretical apparatus of stratigraphy. Simakov does not change the author's assessments of importance of the postulates, he points out only that their abundance reflects the methodological immaturity of the science, the specificity of subjects under study, and heterochronous influence of various cognitive ideas and paradigms on stratigraphy.

As suggested by Meyen (1974, 1989), the fundamental independent principles of Steno and Smith–Huxley along with the chronological interchangeability of characteristics designated as the Meyen principle are taken as operational in stratigraphy; their application is considered as necessary and sufficient to establish the space and time relationships between stratigraphic units (early, late, coeval, concurrent).

Simakov proposed an important addition to the definition of the first principle given by Meyen. The addition refines the relationship between geologic bodies like "lower/higher." The Steno principle implies: "The temporal relationships between geological bodies as earlier/later are predeterminated by their original spatial relationships like lower/higher and (or) by genetic links." I believe that the genetic links, incorporated into the definition and capable, according to formulation ("or"), of replacing the original relationships "lower/higher," should be somehow restricted in use. For the stratigraphic unit obeying the law of superposition, the relationships imply only that its formation (genesis) is associated with gravitation (Zhamoida, 1994).

Meyen did not formulate the second and third principles and had no objections against my formulation. Simakov gives two definitions of the Huxley principle. The first holds that "in homologic sections of paleosystems of any organization level, the similar faunas occupying a similar position are coeval" (p. 103). This strict definition is attributed for some reason to paleontological characteristics only. Huxley, who suggested the term "homotaxis," wrote about geological bodies—the series of layers. Besides, the terms "homologic sections" and "similar position" call for a special explanation, because "homology" and "analogy" in geology is not identical to the same notions in biology, although Huxley considered them as equivalents.

The proposed general formulation of the principle that "homologic stratigraphic units located symmetrically in homologic sequences are coeval" (p. 105) also requires the further clarifying. I prefer the definition which, in my opinion, more closely corresponds to the

[4] It is remarkable that the *Stratigraphic Code of the USSR* (1977) does not suggest criteria of distinguishing the different-rank taxons of the general (international) scale on the basis of first occurrence levels of new paleontological taxons of a certain rank, as it was recommended, for instance, in the brochure *Stratigraphic Classification, Terminology, and Nomenclature* (1965).

operational character of the principle: "the stratigraphic correlation of particular sections (if their direct links cannot be traced) should be based on comparison of homotaxial, of i.e., identical successions of identical characteristics, including environments and events of the past" (Zhamoida, 1984, p. 12). The word "identical" is introduced here in accordance with a good reasoning by Simakov.

In his explanation of the third principle, Simakov distinguishes two aspects: on the one hand, the principle "works" for the correlation of stratigraphic units, and on the other, for tracing their boundaries. It seems to be correct and stressing the universality of the principle. The formulation of the principle (it is termed unexpectedly as "proposition") is suggested from the modern point of view as follows: "The polymorphic consequences of the same event are topologically concurrent" (p. 106). I interpreted differently the operational meaning of the Meyen principle: "Variable, partially overlapping spatial range and integration of stratigraphics ensure their chronological interchangeability, providing grounds for the intra- and interregional (up to global) correlation based on the gravest characteristics" (Zhamoida, 1984, p. 13).

In 1974, when Meyen submitted his monograph substantiating the above fundamental principles of stratigraphy, Sokolov (1974, p. 12) considered the possibility of using the parallel (with respect to standard) biozonations and termed the procedure of boundary tracing as the transmission of the correlation function, which allows the characteristics of zonal species assemblages to be extended within the limits of zonation accuracy. In fact, Sokolov paid attention to the principle of chronostratigraphic interchangeability of characteristics as to the important one.

I consider also as fundamental the principle of the unbiased and unique nature of stratigraphic units, as it is termed by Stepanov and Mesezhnikov (1979), or the Halphin–Stepanov principle in terms of Simakov (p.16). Meyen accepted this principle as well and considered it as closely related to the statement that the stratigraphic procedure is not a classification. He placed it in his list even before the Steno principle (Zhamoida, 1984, 1995). Sokolov also believes that the principle in question is first among "the four main fundamentals of stratigraphy" (1991, p. 8) emphasizing that the stratigraphic unit "is a unit characterizing a *real* geological setting and its natural limits."

The "laws" and "principles" listed by Simakov in Table 1 are denoted in the book as propositions claiming the fundamental status in stratigraphy, but he did not considered their actual affiliation exactly to this science. His attempt to relate these propositions with the fundamental cognitive laws (Table 2, p. 54) deserves attention, but we should "rank" first the propositions in accord with their significance for stratigraphy. Even if somebody extends the scientific sphere of stratigraphy, he must evaluate this significance, although this is not an easy and special feat. The assessment of some "principles" and "laws" has been considered in a series of works by Simakov himself (1989, 1995), and also by L.L. Khalfin, S.V. Meyen, A.M. Sadykov, D.L. Stepanov, M.S. Mesezhnikov, A.I. Zhamoida, I.P. Sharapov, and others.

Stratigraphy and its subject-matter. Simakov gives no definition of stratigraphy, although he points out the important peculiar features of the science. "Stratigraphy mainly deals with the records of the functioning geologic, geobiocoenotic (ecologic), and biologic systems composed of inert, bioinert, and biogenic components" (p. 43), which form the stratigraphic units and the whole hypergene shell of the planet. Geologic systems evolve in a linear manner and multicycles of their evolution partially shade the irreversibility of the latter (Rundkvist, 1990). Geobiocoenotic and biologic systems evolve in the nonlinear mode and indications of their irreversible qualitative transformation conceal the cyclic evolutionary patterns.

Since Simakov takes for stratigraphic unit characteristics only those originated in the course of the unit formation and reflecting the "primary properties and relations of the units" (p. 68), I understand his words "evolution of the system" as "formation of the system." Being open, these systems obey the law of "allometric evolution" (Zhirmunskii and Kuz'min, 1990).

Simakov considers the law of "metachronous evolution" as controlling all three types of the systems. This ensures an objective opportunity of using different methods in stratigraphy in order to distinguish stratigraphic units different in their essence. In biologic systems, an effect of the law is dissimilarity of biostratigraphic units distinguished on the basis of different groups of faunas (floras). The aforesaid predetermines an objective necessity to choose one or another reference scale of the conceptual time, in other words, to choose one of characteristics (features) of geologic or biocoenotic system and one (orthochronological) group in the biologic system. As a result, Simakov remarks that it is "impossible to use the principle of long-distance action in the theoretical cognitive apparatus of geology and the related complex approach to discrimination between stratigraphic units and to substantiation of the unit boundaries in stratigraphy" (p. 49). This conclusion naturally leads him to conclude that the efficiency of "using the past events and action of some global factors as responsible for appearance of "natural" boundaries" (p. 44).

I think that it is a simplified understanding of the event stratigraphy. It should be noted first that all stratigraphic operations (i.e., the stratigraphy as a whole) were and are aimed to reveal the past events. The modern approach to the event stratigraphy assumes the use of a series of methods revealing a set of characteristics of a given specific boundary. Second, the geographic criterion is used to rank the "events," and global events (why should they not exist?) are regarded as the highest

in rank. Third, exactly the synchronous but variable events (their traces) in different areas secure the "action" of the Meyen principle of chronological interchangeability of characteristics.

Thus, the principle of long-distance action seems to be inappropriate for the theoretical cognitive apparatus of geology, but in practice, it means the integration of methods revealing a series of theoretically independent characteristics, which actually secure the performance of three principal tasks in stratigraphy. Stratigraphic units distinguished by different methods and characterizing different systems, when combined, may distinguish a system of a higher category within which they interplay. The seismo- and biostratigraphic units and their boundaries are not indifferent to lithostratigraphic units and their boundaries, are they? The latter are often interrelated with biostratigraphic units and so on. Here, if we take into account the principal or specialized tasks of stratigraphy, my notion of "the relative weights of characteristics" appears to be important.

According to Simakov, the subject matter of stratigraphy "is the spatial–temporal properties and relationships between natural bodies (stratigraphic units)" (p. 11). I believe the Meyen's formulation (1989, p. 51) depicts more accurately the specificity of stratigraphy: "original spatial-temporal relationships between geological bodies and *their features that are indicative of these relationships*" (spacing is mine). The Simonov's expression "spatial–temporal properties" mean is not quite clear.

Among the tasks (better to say, principal tasks), he designates three: the distinguishing of stratigraphic units, their classification (to be more exact, systematization), and the establishing of their spatial–temporal relationships. In my opinion, three principal tasks presented in the introduction to native stratigraphic codes are not only obviously and casually interrelated with three operational principles of stratigraphy, they afford the more complete and logical performance of the ultimate objective, which is formulated by Simakov as "the establishing of regional and global time–space models of stratigraphy" (p. 12). Afterward, however, he closely repeats the tasks specified in our codes but designates them as typical. The indicated typical tasks are, first, *the distinguishing* of stratigraphic units, second, the compilation of *regional schemes* of their original spatial–temporal relationships, and third, the simulation of the primary geological space and time structure of the *global scale*" (p. 58). It is still unclear to me how the principal tasks of stratigraphy differ from typical ones.

In general, I have the impression that Simakov shares the general definition of stratigraphy given by Meyen (1989, p. 51), who considers it is a science studying the "spatial–temporal relationships between rock complexes (geologic bodies) in the Earth's crust." At the plenary session of the Interdepartmental Stratigraphic Committee, which approved the last "Stratigraphic Code" of 1992, the majority supported exactly this formulation. Consequences of the formulation are considered in my work (Zhamoida, 1994).

CONCLUSION

Let me point out principal points of the above discussion.

Simakov has analyzed and explained the most significant notions of theoretical stratigraphy and made a serious attempt to tie his propositions, pretending to the status of stratigraphic laws and principles, to the fundamental cognitive laws in order to prove the objectivity of the former and to define the place of stratigraphy in the family of natural sciences.

He contributed much to the understanding of the systems nature of stratigraphic subjects, explained and gave his own interpretation of such peculiar features of geologic systems as their openness and statistic nature, their stage-by-stage evolution. He also demonstrated the autonomy of evolution laws concerning different systems and processes, and set forth the problems of the linear and nonlinear evolution of the systems.

The most outstanding result of his study is a guide to creating the metrics of the conceptual geologic time, beginning with definitions of such stratigraphic notions as the age, duration (continuance), succession, incorporation, and synchronism. In addition, peculiarities of geologic time, such as its condensed state, anisotropy, and nonuniformity are considered together with proposals for solving the problem of measuring the static time.

In addition, many other problems are dealt with in the work to some extent. Not all the questions and problems presented are solved of course, as it is equitably pointed out by the author.

A further elaboration of notions concerning the linearity and nonlinearity of evolution in stratigraphy, as well as in geology, is required. The problem of relations between antagonistic features of geosystems, such as their openness and independence, remains unsolved and leads to the inefficient use of the Meyen principle in both the theory and practice.

Lending a little optimism is Simakov's conclusion, when he regards the ways of solving the problem of measuring the geologic time and shows that it depends on the solution of the classification problem. In his opinion, the latter is not only unsolved, but even the principles of taxonomy are not elaborated.

I would like to emphasize the importance of the fact that relationships between natural geologic bodies, such as the stratigraphic unit, formation, and facies, are poorly understood, like the problem of stratigraphic unit boundaries in their lateral boundaries (of local and regional categories) and vertical successions. Certainly, stratigraphy should use achievements of other sciences, including nongeological ones, but *for its own purposes* and for performance *of its own tasks*, namely for understanding the space–time pattern for all geologic disci-

plines, as was rightly noted by Simakov. The general terms like litho-, bio-, magnetostratigraphy, and so on imply only the application of corresponding methods in stratigraphy but not the independence of these branches in the science. Methods of various sciences must be used in stratigraphy, but this does not mean that, for instance, "stratigraphy may turn into the formational analysis at all levels from local and regional to international and global" (Shvanov, 1996, p. 26).

The basic objective of all stratigraphic investigations is to reveal and trace with their help the coeval or synchronous records of past geological events and phenomena, which is equally necessary for the geologic mapping and studies on facies and formations, as well as for mineral prospecting and ecological studies from retrospective to prognostic. It is not by chance that organizing the International Geological Correlation Program (IGCP) the first Working Group combined all the available stratigraphic programs. Thus, from none standpoint, stratigraphy "is if not a hostage, but at least a junior partner of the facies and formation analysis" (p. 8).

Simakov suggests the lines of attack the problems of theoretical stratigraphy as follows (p. 8):

(1) elaboration of the general methodological approach to the analysis of structure, genesis, and evolution of subjects of stratigraphy and allied disciplines (first of all, of lithology, facies and formation analysis) with due regard to the system approach;

(2) algorithmic presentation of the procedure of distinguishing subjects of stratigraphic studies and units mentioned above (I believe we should combine here the distinguishing and correlation tasks);

(3) development of classification of stratigraphic subjects, which is closely related to and depends on the progress in classifications of formations and facies;

(4) solution of fundamental problems concerning the theory of time measurements, the static and, first of all, the geological time included.

One cannot disagree with the necessity of solving these problems. However, I cannot escape a sad impression reading the list of problems. It means that nothing, or almost nothing, is done to develop the cognitive aspect of theoretical stratigraphy, and everything must begin at the beginning. Is it so? I believe not. Even the works of our compatriots only, the book under discussion included, point to a different situation, though the way to establish an adequate theory of stratigraphy will certainly be hard and long.

Reviewer A.Yu. Rozanov

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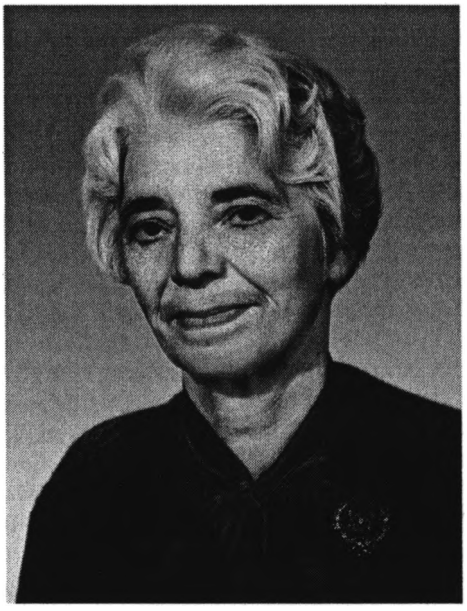
In Commemoration of Viveya Pavlovna Vasilenko (1915–1998)



Viveya Pavlovna Vasilenko, Doctor of Geology and Mineralogy, a distinguished specialist in studying foraminifers and Cretaceous biostratigraphy of Russia, passed away in her eighty-third year on March, 14, 1998, in St. Petersburg.

Vasilenko was born on September 6, 1915, in Tiflis. As a young girl, she lived in Vladikavkaz in the large and friendly family of an architect Pavel Pavlovich Schmidt.

Her father graduated from the St. Petersburg Institute of Civil Engineers. He contributed much to the architectural development of Vladikavkaz, where he lived and worked from 1904 to 1942. His wife, Erminiya Frantsevna Proshinskaya was a foreign language teacher, good musician, and enthusiast of an amateur theatre. She was born in a family that settled in Vladikavkaz when it was the Fortress Kapkai. Vasilenko cherished the memory of the family home all her life. With a warm heart, she imbibed a spirit of art and mutual assistance, love and respect for labour, a steady interest to the history of her family and native land, as well as the lasting cultural traditions.

After finishing the secondary school in 1931, Viveya Schmidt left for Leningrad, the city that won her heart in youth for her whole life. Her primary impression of Leningrad was associated with the State Public Library, where she worked in 1932–1933, and with the Bayanus' Courses of Foreign Languages. In 1933, she entered the Leningrad State University and graduated from the Faculty of Soil Geology in 1938. In August of the same year, she joined the Micropaleontological Laboratory at the Oil Geological Prospecting Institute (later VNIGRI) to work there for a half a century. The beginning of her scientific life coincided luckily with the commencement of micropaleontological investigations important for oil prospecting in the USSR. Vasilenko happened to be a researcher at the first and preeminent native micropaleontological laboratory, established in 1930 by A.V. Fursenko, the distinguished expert in protistology. Among her senior colleagues, who were engaged in the study on foraminifers, there were L.G. Dain, N.N. Dampel', N.A. Voloshinova, N.N. Subbotina, E.V. Myatlyuk.

The excellent university background, innate leaning toward paleontology, knowledge of foreign languages, and devotion to work allowed Vasilenko to run quickly through the primary micropaleontological training and to get into the action at the laboratory. As early as 1940, she wrote a paper on the Jurassic and Early Cretaceous foraminifers and stratigraphy of the Cape Il'ya (the Khatanga Bay of the Laptev Sea). Issued only in 1951, her work, along with the paper by Myatlyuk (1939) on Jurassic foraminifers from the same region, laid the foundation of micropaleontological studies on the Mesozoic of the Arctic regions.

During the war years, when she was in Guriev as an evacuee, Vasilenko joined the project of micropaleontological studies on Upper Cretaceous deposits in the

Caspian lowland. A new scheme of the subdivision of Upper Cretaceous deposits in the South Emba oil-bearing region was elaborated and based on foraminifers by her and Myatlyuk in 1943 (published in 1947). They distinguished index foraminiferal assemblages typical of most stages and substages, which became important for the applied geology and fundamentally valuable for subsequent stratigraphic analysis of foraminifers from the Russian platform and western Kazakhstan.

After the war, Vasilenko carried out studies in the Caspian lowland and Mangyshlak, Volga region, Dnieper–Donetsk depression, and Donbas (Donetsk basin). The Paleocene foraminiferal assemblage of the Dnieper–Donetsk depression, originally described by her and correlated with the synchronous fauna of Sweden, was detected later in other southern regions of the USSR and in western Siberia; these results caused great scientific and practical interest. To this day, the work by Vasilenko and Art. Miklukho-Maklai (1964), which is based on the fossil collection from the Usinsk–Rogovsk interfluve area, remains the only paper published on the Late Cretaceous foraminifers from the polar area adjacent the Urals from the west.

In addition to her research work, Vasilenko always was busy with routine paleontological determinations, usually urgent and often lying beyond the scientific sphere. Her results of fauna determinations in foreign samples (in Miocene rocks from Albania and in Cretaceous flysch deposits of Sri Lanka studied in the 1950s and 1970s) remain unknown to the scientific community, being unpublished so far.

Her monograph *Anomalinids* published in a series *Fossil Foraminifers of the USSR* (1954) Vasilenko was defended as the Cand. Sci. dissertation in 1958. This fundamental synthesis on taxonomy of Anomalinidae, one of the foraminifer families significant for the boreal Meso–Cenozoic stratigraphy, included a description of 120 species and the table of their vertical ranges. This work was used for years as a reference book for specialists in micropaleontology in the whole USSR. Contribution of Vasilenko to the academic issue *Fundamentals of Paleontology*, the volume *Protozoa* of 1959, illustrates a great scientific value of her research work.

Studies on foraminifers and stratigraphy of Cretaceous deposits in the Mangyshlak occupy an important place in the scientific biography of Vasilenko. In 1948, she was the first to study the Late Cretaceous foraminifers from natural exposures in the North Aktau Range. Her pioneering microfaunal method used to study the sequence of white chalk beds resulted in recognition of all stages and substages of the Upper Cretaceous, and also in the determination of the units' boundaries, and even in a more detailed subdivision of some units.

The classic work by Vasilenko *Foraminifers from Upper Cretaceous Deposits of the Mangyshlak Peninsula (Description, Schemes, Phylogenesis of Some Groups and Stratigraphic Analysis)* was published in 1961, and in 1964, she received the academic degree of Doctor of Geology and Mineralogy for this work. The book presented a description of 150 species of foraminifers with splendid illustrations made by artist N.A. Ipatovtsev and became a valuable guide to the Late Cretaceous fauna. Very definite, strict, concise, and simultaneously comprehensive descriptions place the book among the highest achievements of Russian micropaleontology. The suggested foraminiferal zonation was a starting example for other stratigraphic schemes based on foraminifers from the Russian platform, Central Asia, Caucasus, and Carpathians.

In the late 1950s, Vasilenko solved another problem concerning foraminifers from the Lower Cretaceous terrigenous, often carbonate-free deposits. Already at the first stage of her investigation, she correlated data on foraminifers with the detailed stratigraphic scheme based on ammonites by Saveliev (1963). However, it was the long-standing scrupulous work to turn the mosaic picture of laterally variable foraminiferal assemblages into the stratigraphic zonation of Lower Cretaceous deposits in the Mangyshlak. The results of the work are included in *Atlas of Characteristic Foraminifers from Lower Cretaceous Deposits of the Caspian Lowland, Mangyshlak Peninsula, and Ustyurt* of 1988, which was published in collaboration with Myatlyuk.

The atlas presenting a description of 248 species and data on vertical and geographic ranges of more than 400 taxa of foraminifers is indeed an encyclopedia of the Lower Cretaceous foraminiferal faunas from terrigenous deposits. The preparation of the atlas for publication was a feat of scientific heroism, as both authors were of an advanced age. It was the last work of the first generation of scientists who worked at the VNIGRI Laboratory of Microfauna.

Vasilenko will endure in the history of Russian paleontology as the author of three monographs and a series of fundamental papers representing an era in the study on foraminifers and biostratigraphy of the Cretaceous deposits in the southern USSR. Her contribution to the unified stratigraphic schemes of the Russian platform and western Kazakhstan is great. Being a curator of the Commission on Micropaleontology at the USSR Academy of Sciences and a member of the Cretaceous Commission of the ISC, Vasilenko consulted native and foreign specialists, expended much time in supervising works of post-graduate students and applicants for academic degrees. For her contribution to the development of the native science and oil prospecting studies, she was decorated with five medals and bestowed with many other awards.

The scientific biography of Vasilenko is not rich in bright events; the every-day working life prevailed. Her 50-year professional life elapsed in the VNIGRI Laboratory of Microfauna with the entrained, concentrated work at binocular microscope. Vasilenko was inextricably linked with the laboratory, she was its flesh of the flesh. She grew up as a specialist there, her works

enhanced the glory of the laboratory, her personality sublimated the moral climate in our collective. The self-respect, high perfectionism, and sense of justice were characteristic of Vasilenko. She regarded works of her colleagues with professional interest and was always eager to emphasize their contribution to her own works. We recall her gratitude to elder colleagues, especially to Fursenko and Myatlyuk. She highly esteemed A.A.Gerke the science editor of her monograph, showed an immutable delight with masterly "dotted" drawings by Ipatovtsev, and appreciated her husband V.K. Vasilenko for his advice of professional stratigrapher and unfailing moral support.

Vasilenko was an affectionate and devoted wife, who almost left the fascinating work for several years in order to brighten up and ease the last period of her husband's life. The sense of duty, compassion, and readiness to help people were qualities inherent in her, and well known to her relatives, friends and colleagues, or simply to neighbours.

Vasilenko was a beautiful, jovial woman, who loved and knew how to dress well despite a possible deficiency in the family budget. She was a good hostess, loving and wise mother, grandmother, and aunt, being the attraction center for the kindred clan. Visits to the Philharmonic Society, museums, and art exhibitions, walks through the city and suburbs, as well as a keen interest in history, were important to her. At the same time, she was an arid mushroom seeker and took up gardening with great enthusiasm.

Viveya Pavlovna Vasilenko will be forever in the hearts of all colleagues, who knew her as an outstanding expert in paleontology, and a noble and versatile gifted person preserving traditions of many generations of her ancestors, the Russian intellectuals.

V.P. Vasilenko is buried in the Northern Cemetery of St. Petersburg.

A.Ya. Azbel', L.V. Vasilenko, E.M. Kaplan, and B.S. Sokolov

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